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1 The role of melting on the geochemical evolution and  
2 isotopic variability of an anatectic complex in the Iberian  
3 Variscides

4

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16

17 Abstract: Formation and evolution of migmatite-granite terranes usually involve several  
18 complex petrological and geochemical processes that leave their imprint on elemental  
19 and isotopic signatures. In this paper, we assess the role of melting reactions on the  
20 genesis of the Variscan Figueira de Castelo Rodrigo-Lumbrals Anatectic Complex  
21 (FCR-LAC) at the Central Iberian Zone, culminating in the generation of abundant S-  
22 type granites. At odds with the proposed for several anatectic complexes elsewhere, it  
23 is demonstrated that no isotopic (Sr-Nd-Hf-Pb) disequilibrium occurred in the FCR-  
24 LAC, in the transitions from metatexites to diatexites and, finally, to granites. However,

25 variable contribution of the source minerals generated batches of melts characterized  
26 by distinct parent/daughter ratios, which explain the significant heterogeneity of  
27 present-day isotopic signatures. While Rb/Sr ratios and the Pb budget of the  
28 successively generated lithotypes were mainly controlled by major minerals, such as  
29 muscovite, K-feldspar and plagioclase, the accessory phases played a major control on  
30 Sm/Nd and Lu/Hf ratios. Our study demonstrates the existence of two distinct diatexite  
31 groups produced by different reactions. Type-1 diatexites, having high Rb contents and  
32 fractionated HREE, are geochemically more akin to the associated S-type granites.  
33 They were produced via fluid-absent reactions during dehydration-melting of muscovite  
34 with production of peritectic K-feldspar, sillimanite and melt. The less abundant type-2  
35 diatexites required influx of external fluids during melting reactions, which consumed  
36 more plagioclase than muscovite.

37 Isotopic data reveal the existence of two groups with distinct initial compositions ( $\epsilon\text{Nd}_{320}$   
38 = -5.05 to -6.03 and  $\epsilon\text{Hf}_{320}$  = -3.42 to -4.45 vs.  $\epsilon\text{Nd}_{320}$  = -7.30 to -8.89 and  $\epsilon\text{Hf}_{320}$  = -6.45  
39 to -8.47), both composed of metatexites, diatexites and granites, which is explained by  
40 source heterogeneity. The main source of the anatectic complex was the  
41 metasedimentary Neoproterozoic – Lower Cambrian Douro-Beiras Supergroup, with  
42 minor contribution of the Ordovician Ollo de Sapo magmatic rocks.

43

44 Keywords: Anatectic complex; Melting reactions; Isotope geochemistry; Variscan  
45 Orogeny

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47

48

49 1. Introduction

50 Late-stage mountain building is typically characterized by crustal anatexis and  
51 generation of granitic bodies, a common feature in the continental tectonic framework

52 since the Archean (Whitney et al., 2004). Upper crust emplacement of granitic magmas  
53 leaving refractory residues leads to intracrustal differentiation, which explains the  
54 compositional distinction between lower and upper crust (Sawyer et al., 2011).

55 The production of large volumes of melt through partial melting of a source rock  
56 depends on protolith fertility, on the presence of fluids, and on the temperature-  
57 pressure conditions, namely the magnitude of the thermal anomaly (Brown, 2013;  
58 Clemens, 2006). Metapelitic and metapsammitic rocks containing large amounts of  
59 hydrous phases, such as muscovite and biotite (30% to 50%), are fertile protoliths and  
60 potential sources of peraluminous melts through metamorphic incongruent melting  
61 reactions (fluid-present or fluid-absent) at 700 – 800 °C (Brown, 2013; Sawyer et al.,  
62 2011). Indeed, fluid-absent melting of micas in metapelites and metagreywackes can  
63 yield up to 50 vol% of melt (Clemens and Vielzeuf, 1987; Sawyer et al., 2011; Bento  
64 dos Santos et al., 2011b; Brown, 2013). For these reasons, migmatite-granite  
65 complexes have been the target of several studies with the objective of exploring links  
66 between high-grade metamorphism, partial melting processes, the origin of the related  
67 granitic bodies and crustal differentiation (e.g. Johannes et al., 2003; Bento dos Santos  
68 et al., 2011a; Brown et al., 2016; Sola et al., 2013).

69 In the Iberian Variscan Belt (the southwestern sector of the European Variscan  
70 orogen), more precisely in the Central Iberian Zone (CIZ), several authors have  
71 emphasized the relationship between some granites and migmatitic rocks (Areias et al.,  
72 2014; Carrington da Costa and Teixeira, 1957; Ferreira et al., 2014; Pereira et al.,  
73 2017; Ribeiro et al., 2011; Vanderhaeghe, 2009). The Figueira de Castelo Rodrigo-  
74 Lumbrales Anatectic Complex (FCR-LAC) is one of such anatectic complexes and has  
75 the particularity of showing a clear spatial relationship between metatexites, diatexites  
76 and granites. In this way, this complex offers the opportunity to assess the melting  
77 reactions that genetically link migmatites (metatexites and diatexites), granites, and  
78 their metasedimentary sources. Links between the FCR-LAC granites and the hosting  
79 pre-Ordovician metasedimentary rocks have already been suggested by Ferreira et al.

80 (2019), based on the study of inherited zircon grains, a question that will be  
81 readdressed in this paper.

82 Radiogenic isotopes are widely applied to identify source characteristics of  
83 magmatic rocks (e.g. White, 2010). Indeed, it is usually assumed that the isotopic  
84 composition of a magma is similar to that of the source rock, a hypothesis implying that  
85 no isotopic fractionation occurs during melting and that almost the same happens with  
86 the parent (P)/daughter (D) element pairs. This is true if each of the source mineral  
87 phases is not a significant repository of any of the P or D elements for the isotopic  
88 systems considered, i.e. when the P and D elements are highly incompatible. However,  
89 in silica-rich magmatic systems, such as S-type granites and related anatectic rocks,  
90 isotopic disequilibrium becomes more common (e.g. Himalayan leucogranites - Ayres  
91 and Harris, 1997; Anatectic Complex of Toledo - Barbero et al., 1995; Sierra Nevada  
92 Batholith - Tommasini and Davies, 1997; Southern Sierra Nevada - Zeng et al., 2005b;  
93 see also Wolf et al., 2019;). Indeed, residue/melt isotopic disequilibrium can take place  
94 due to a) insufficient temperature to reach the closure temperature of some of the  
95 accessory phases, with which P and/or D elements are compatible, resulting in a  
96 significant fractionation (up to  $> 20 \text{ } \epsilon_{\text{Hf}}$  units; Tang et al., 2014); and b) chemical  
97 diffusivity being sluggish compared with the time frame necessary for melt extraction  
98 from the residue (Ayres and Harris, 1997; Farina and Stevens, 2011; Zeng et al., 2005;  
99 Tang et al., 2014; Wolf et al., 2019). These factors can lead to the perception that  
100 isotopic disequilibrium during anatexis is ubiquitous (e.g. Tommasini and Davies,  
101 1997). However, in some cases, even for situations specifically investigated for isotopic  
102 disequilibrium, no significant evidence—was found (e.g. Wolf et al., 2019). In this  
103 perspective, the isotopic variability of an anatectic complex must also be addressed in  
104 light of regional source heterogeneities (e.g. Yakymchuk et al., 2015, 2013), and on the  
105 role of melting reactions leading to distinct P/D pairs with the consequent development  
106 of distinct isotopic signatures due to variable rates of radiogenic ingrowth through time.

107 For these reasons, the importance of isotopic disequilibrium during anatexis is still  
108 inconclusive.

109 Distinct isotopic systems are expected to behave differently during anatexis, with  
110 some of them responding variably to diverse melting conditions (e.g. Wolf et al., 2019).  
111 This confers to multi-isotopic studies the potential to investigate the hypothetical role of  
112 isotopic disequilibrium and source heterogeneity, as well as to better decipher the role  
113 of distinct melting reactions during the different stages of anatexis.

114 Therefore, in this study, we use Sr, Nd, Hf and Pb isotopes, the elemental  
115 whole-rock compositions, and zircon and apatite trace-element compositions from the  
116 FCR-LAC lithologies to: 1) assess the melting reactions behind the different lithotypes;  
117 2) evaluate the potential role of isotopic disequilibrium during anatexis; 3) provide  
118 insights onto the causes of isotopic variability characterizing the anatectic complex; and  
119 4) explore the geochemical link between the different stages of anatexis, from  
120 migmatites to granites, and also to infer potential metapelitic protoliths.

121

## 122 2. Geological Setting

123 The European Variscan Belt is the result of a complex evolution culminating with the  
124 polyphasic collision of the continents Laurentia and Gondwana during the Devonian –  
125 Carboniferous periods (e.g. Dias and Ribeiro, 1995; Nance et al., 2010). In the internal  
126 zones of the orogen, the continental collision produced the rapid formation of  
127 metamorphic core complexes, characterised by the exhumation of migmatites, large  
128 volumes of granitic magmas, and LP-HT metamorphism (Burg et al., 1994; Schulmann  
129 et al., 2002).

130 The FCR-LAC is located within the autochthonous terranes of the variscan Central  
131 Iberian Zone (CIZ) and it is an example of a migmatite-granite complex formed during  
132 the Variscan Orogeny (e.g. Ferreira et al., 2019) (Fig. 1). This anatectic complex  
133 preserves significant field macroscopic evidence of the gradual anatectic evolution from

134 metatexites, to diatexites and to syn-tectonic granites (Fig. 2). The FCR-LAC contacts  
135 with low-grade (biotite and chlorite zone) metamorphic units of Ediacaran-Cambrian (to  
136 the North and South) and Ordovician age (to the South) through the Huebra and the  
137 Juzbado-Penalva do Castelo shear zones, respectively. Their sinistral kinematics were  
138 active at least during the intracontinental collision stage ( $D_3$ , third phase of the Iberian  
139 variscan deformation; Pereira et al., 2017 and references therein), although it has been  
140 suggested that it represents a much older tectonic discontinuity (Iglesias and Ribeiro,  
141 1981). The Juzbado-Penalva do Castelo shear zone (JPCSZ) is a 200 km long, 5 to 15  
142 km wide first-order structure, with a proposed horizontal displacement between 65 and  
143 100 km (Iglesias and Ribeiro, 1981; Villar Alonso et al., 2000). Geothermobarometric  
144 estimates using mineral equilibria point to peak metamorphic conditions of  $T = 761 \pm 50$   
145  $^{\circ}\text{C}$  and  $P = 5.0 \pm 1.0$  kbar (Pereira et al., 2017). Exhumation mechanisms of this  
146 anatectic complex have been widely discussed and two major models have been  
147 proposed. Díez Fernández and Pereira (2016) invoked extensional tectonics and  
148 orogenic collapse as the mechanism responsible for anatexis and upper crust  
149 emplacement, followed by capture by the strike-slip shear zone. However, the  
150 interpretation of the complete exhumation process of this granitic-migmatitic  
151 association must consider that this complex occurs in contact with low grade  
152 metamorphic rocks to which is juxtaposed by first-order high-angle shear zones, also  
153 evidenced by recent magnetotelluric studies (Alves Ribeiro et al., 2017). Migmatites  
154 and the associated granites preserve a low dipping, non-horizontal transport lineation  
155 ( $6^{\circ}$ – $12^{\circ}$ ; Pereira et al., 2017), clearly indicative of a significant net vertical mass  
156 transfer during the 65 – 100 km horizontal displacement of the JPCSZ, caused by  
157 simple shear-dominated transpression during the  $D_3$  stage (see Pereira et al., 2017).  
158 Petrochronological constrains recently published estimated cooling rates as high as 35  
159  $^{\circ}\text{C}.\text{Ma}^{-1}$  and fast exhumation rates (max of  $0.84 \text{ mm}.\text{a}^{-1}$ ), which confirm a tectonically-  
160 assisted exhumation process (Ferreira et al., 2019).

161

162 3. Field and petrographic observations

163 The low-grade (chlorite-biotite zones) metamorphic units are composed of phyllites  
164 and quartzphyllites, affected by a nearly E-W schistosity (Fig. 3a). Phyllites in the  
165 vicinity of granites show post-deformation andalusite porphyroblast growth, usually  
166 pseudomorphosed, formed by contact metamorphism (Fig. 2a and Fig. 3b). The  
167 mineral assemblage of these phyllites also includes quartz, muscovite, biotite, and  
168 minor plagioclase and chlorite, as well as accessory titanite, zircon, apatite, rare  
169 staurolite, and opaque minerals.

170 Metatexites exhibit stromatic structures still preserving a pre-melting banded  
171 orientation inherited from the regional E-W schistosity (i.e. inherited fabric). Sometimes,  
172 centimetric or millimetric lenses of peritectic sillimanite associated to muscovite are  
173 found (Fig. 2b). Occasionally, metatexites are intersected by centimetric leucosome,  
174 sometimes boudinated, and filled with granitic to pegmatitic material (Fig. 2c).  
175 Leucosome patches up to 20-35 cm, also occur in association with these veins (Fig.  
176 2d). The large majority of leucosomes are concordant to the pre-migmatization  
177 structures (schistosity) of the hosting rocks. The melanosome of the foliated metatexite  
178 is predominantly composed of biotite and muscovite. In contrast, leucosomes are  
179 predominantly composed of a quartz-feldspar assemblage (Fig. 3c). Fibrolite appears  
180 as relic grains in late muscovite formed during retrogression (Fig. 3d). The most  
181 common accessory phases are zircon, apatite, tourmaline, staurolite and ilmenite.

182 Diatexites show structures such as restitic nodules (Fig. 2e), schlieren (Fig. 2f), and,  
183 occasionally, ptygmatic folding. Closer to the shear zones, diatexites reveal E-W  
184 oriented shear deformation planes, i.e. the same orientation as the shear zone. Locally,  
185 there are pegmatitic and leucosome veins forming vein-structured diatexites. Some of  
186 these diatexites show a near-nebulitic texture (only slightly more coarser-grained and  
187 biotite-enriched than the examples described in Sawyer (2008)). As a whole, diatexites  
188 comprise quartz, plagioclase, K-feldspar, biotite, minor fibrolitic sillimanite and



189 secondary muscovite (Fig. 3f). Chemically, two different groups of diatexites can be  
190 distinguished (see section 6.2), which are not macroscopically distinguishable.  
191 However, thin section petrography reveals that the mineralogical composition is slightly  
192 different, with type-1 diatexites having more fibrolitic sillimanite and secondary  
193 muscovite, but lower amounts of K-feldspar than type-2 diatexites (Table 1).

194 Additional evidence of partial melting in these migmatites can be found in the  
195 occurrence of corroded grains of biotite and melt films (Fig. 3e), mainly in quartz-  
196 feldspar boundaries.

197 In general, the contact between the metatexites, diatexites and granites is marked  
198 by a gradual transition, but there are also injections of granitic material found in the  
199 migmatites, feeding the S-type plutons.

200 Granites are essentially two mica-bearing, with biotite and muscovite appearing in  
201 different proportions and with variable grain sizes. These variations led to their  
202 grouping onto 10 different granite facies (Silva and Ribeiro, 2000; Fig. 2g and Fig. 2h).  
203 Some granitic outcrops reveal deformation structures like those found in the  
204 migmatites, compatible with the JPCSZ movement (Fig. 3h). These granites are mainly  
205 composed of quartz, plagioclase, K-feldspar, biotite, muscovite (Fig. 3g) and minor  
206 fibrolitic sillimanite. Accessory mineral assemblage includes zircon, apatite, rutile  
207 needles and ilmenite.

208 Apatite from these granites yield 288 – 307 Ma, while newly grown zircon grains  
209 provide magmatic U-Pb crystallisation ages between 300 and 317 Ma (Ferreira et al.,  
210 2019). However, a significant part of the zircon population is inherited (54%), yielding  
211 ages in the range 400 – 650 Ma, having survived the melting and segregation stages.

212

#### 213 4. Analytical methods

214 We report 55 analyses of whole-rock geochemistry of granites, migmatites and  
215 phyllites from the Douro-Beiras Supergroup that were conducted in the Activation Lab

216 in Ontario, Canada. Major elements were measured by Fusion Inductively Coupled  
217 Plasma Optical Emission Spectrometry and trace elements by Fusion Inductively  
218 Coupled Plasma Mass Spectrometry (see details in the Supplementary Material 1).

219 Whole-rock isotopic analyses of Sr, Nd, Hf and Pb were performed in 13 samples,  
220 including 8 migmatites (5 metatexites, 2 type-1 diatexites and 1 type-2 diatexite), 5  
221 granites and 2 metasedimentary units from the Douro-Beiras Supergroup at the  
222 Laboratoire G-Time of the Université Libre de Bruxelles (ULB, Belgium) using a Nu  
223 Plasma I Multi-Collector Inductively Coupled Mass Spectrometer (MC-ICP-MS)  
224 instrument (for more details see the Supplementary Material 1).

225 Zircon and apatite grains separated from the anatexitic units were analysed for their  
226 trace element content using an ASI RESOLUTION 193 nm ArF excimer laser coupled to  
227 the ANALYTIK JENA Plasma Quant Elite quadrupole ICP-MS at the University of  
228 Portsmouth (UoP) (see more analytical details in the Supplementary Material 1).

229

## 230 5. Results

### 231 5.1. Whole-rock elemental geochemistry

232 Whole-rock elemental compositions are presented in Table 2, where the studied  
233 samples are grouped as phyllites, metatexites, diatexites and granites.

234 Phyllites and metatexites are enriched in  $\text{Al}_2\text{O}_3$ ,  $\text{FeO}^t$ ,  $\text{MgO}$  and  $\text{TiO}_2$  in comparison  
235 to diatexites and granites, evidencing a negative correlation with silica (Fig. 4). This  
236 reflects the larger abundance of refractory minerals (e.g. biotite and titanite) in the  
237 metasedimentary protoliths and in the metatexite's melanosomes, representative of an  
238 incipient degree of partial melting. On the other hand, diatexites and granites  
239 commonly yield higher contents of  $\text{SiO}_2$ ,  $\text{Na}_2\text{O}$ ,  $\text{K}_2\text{O}$  and  $\text{P}_2\text{O}_5$  when compared to  
240 metatexites. This difference is consistent with higher degree of partial melting and,  
241 eventually the occurrence of some crystal fractionation (e.g. Sawyer, 2008), and, thus,  
242 with the production of quartz-plagioclase-, K-feldspar- and apatite-rich melts. Two types

243 of diatexites were identified mainly based on their composition (Fig. 4): type-1  
244 diatexites ( $\text{SiO}_2 = 68 - 74 \text{ wt}\%$ ) and type-2 diatexites ( $\text{SiO}_2 = 74 - 75 \text{ wt}\%$ ).  
245 Additionally, type-2 diatexites are distinguishable by a wider variation of  $\text{Na}_2\text{O}$   
246 concentration, slightly lower average  $\text{Fe}_2\text{O}_3^{\text{t}}$  content and lower concentrations of  $\text{P}_2\text{O}_5$ ,  
247 which are similar to the depicted by metatexites and phyllites.

248 All lithotypes display high  $\text{Al}_2\text{O}_3$  content ( $\text{Al}_2\text{O}_3 > 12 \text{ wt}\%$ ), being strongly  
249 peraluminous (phyllites:  $\text{ASI} = 2.0 - 3.3$ , quartzphyllites:  $\text{ASI} = 1.2 - 2.4$ , metatexites:  
250  $\text{ASI} = 1.5 - 2.7$ , diatexites:  $\text{ASI} = 1.1 - 1.5$ , and granites:  $\text{ASI} = 1.1 - 1.4$ ).

251 The resemblance between metatexites and phyllites, revealed by major elements, is  
252 also generally depicted by trace elements (Fig. 5). Type-2 diatexites exhibit lower Rb  
253 contents compared to the other anatectic rocks, but similar to metasediments. Phyllites,  
254 quartzphyllites, and metatexites have significantly similar patterns in the multi-element  
255 diagrams, showing negative anomalies of Nb, K, Pb, Sr and Ti, and positive anomalies  
256 of U, Ta, La, Ce and Nd (Fig. 6a). Zr and Hf positive anomalies are evident in the  
257 quartzphyllites (Fig. 6a), as the result of their higher zircon contents (observed under  
258 the microscope), in agreement with their more psammitic characteristics in comparison  
259 with the remaining metasediments. Multi-element patterns of diatexites and granites  
260 overlap within error (Fig. 6b), exhibiting a wide range of values for some elements,  
261 such as Th, REE and Zr-Hf. Both lithotypes show negative anomalies for Nb, Pb, Sr  
262 and Ti. There are some similarities between metasediments/metatexites and  
263 diatexites/granites, as all yield Nb, Pb, Sr and Ti anomalies (Fig. 6a and 6b).

264 REE composition data are summarized in Table 2 and Figure 7. Phyllites,  
265 quartzphyllites and metatexites have almost identical normalised patterns (Fig. 7a),  
266 with all but one sample being included in a very limited concentration range (e.g. La =  
267 41 to 61 ppm; Yb = 2.2 to 3.5 ppm). The exception is one metatexite sample (JTJ-60A)  
268 which stands out of the group, because of its lower REE contents (La = 18; Yb = 1.4),  
269 but yet yielding similar  $(\text{La}/\text{Yb})_{\text{N}}$  ratios (7.43 to 13.35). The Heavy Rare Earth Elements  
270 (HREE) are weakly fractionated ( $\text{Dy}/\text{Yb}_{\text{N}} = 1.07 - 1.35$ ) and the Eu anomalies are small

271 (Eu/Eu\* = 0.5 – 0.7). Diatexites and granites display a wide range of REE  
272 concentrations ( $\Sigma$  = 8 – 777 ppm) (Fig. 7b and 7c), despite the studied granites having  
273 similar normalised patterns across the different facies. At odds, diatexites have a wider  
274 variability, showing different degrees of HREE fractionation ( $Dy/Yb_N$  = 0.8 to 5) and Eu  
275 anomalies, either negative (down to 0.2; type-1 diatexites) or positive (up to 4; type-2  
276 diatexites). See section 6.2 for more details on the chemical differences between these  
277 two types of diatexites.

278

## 279 5.2. Accessory minerals: trace element composition

280

### 281 5.2.1. Apatite

282 The trace element compositions of apatite grains from the FCR-LAC and their host  
283 rocks (phyllites) are remarkably distinct in Sr, Y, Th and REE compositions (Table 3;  
284 Fig. 8a-8e). The highest Sr content in apatite was found in phyllites (317 – 1471 ppm),  
285 decreasing gradually in metatexites (113 – 487 ppm), diatexites (86 – 144 ppm) and  
286 granites (74 – 99 ppm). Thorium behaves similarly. On the other hand, Y contents in  
287 apatite are low in phyllites (480 – 854 ppm), are quite similar between metatexites (837  
288 – 1973 ppm) and diatexites (847 – 2026 ppm) and show a large range in granites (212  
289 – 1546 ppm). Concerning their U composition, phyllites reveal lower values (8 – 34  
290 ppm), while metatexites exhibit a larger variation (17 – 223 ppm), identical to that of  
291 diatexites (33 and 185 ppm). The maximum U concentration in the diatexites  
292 corresponds to a type-2 diatexite. The variation of U in granites is smaller, but within  
293 the same range of values as the migmatites (49 and 81 ppm).

294 Apatite grains are highly enriched in REE with  $\Sigma$ REE ranging from 1397 to 5829.  
295 They all show a marked enrichment in LREE compared to HREE, as depicted by  
296 (La/Yb)<sub>n</sub> ratios (Fig. 8f; phyllites = 5.1 – 32.6, metatexites = 0.6 – 3.7, diatexites = 0.6 –  
297 6.8 and granites = 3.7 – 19.1, yet (La/Sm)<sub>n</sub> ratios are lower than the unity, except in Ily

298 granite. At odds, apatite (Tb/Lu)<sub>n</sub> ratios are clearly higher than 1 for all the lithotypes.  
299 The Ily granite has a REE fractionation higher than the remaining granite facies,  
300 because of its higher LREE contents. In general, Eu anomalies are negative with the  
301 exception of the late-tectonic granite (granite Ily: 1.06), which lacks an anomaly.

302

### 303 5.2.2. Zircon

304 Contents of Hf, Y, U, Th and HREE in zircon grains exhibit positive correlations with  
305 those from the whole rock, indicating that zircon controls the budget of these elements  
306 in migmatites and granites (Table 4; Fig. 13b and Fig. 13c). Hf contents are similar  
307 between metatexites (8180 – 14410 ppm), diatexites (11060 – 12750 ppm) and  
308 granites (10280 – 12670 ppm). Lu values decrease in zircon grains from metatexites  
309 (48 – 309 ppm), to diatexites (48 – 91 ppm), and to granites (14 – 58 ppm). We were  
310 only able to determine Y contents for half the measured points (11 out of 23). For the  
311 available measurements, metatexites yield 120 – 2074 ppm, diatexites 610 ppm, and  
312 granites range between 469 – 1199 ppm. U and Th contents are highly variable, with  
313 Th/U ratios in zircon from diatexites (0.06 – 0.11) and granites (0.11 – 5.88) being  
314 broadly higher than 0.1 (10 spots out of 12), whereas metatexites yield a wide range  
315 (0.003 – 0.845), with half the analysed grains showing ratios lower than 0.1 (6 of the 11  
316 spots).

317 Zircon from the different units exhibit high and similar  $\Sigma$ HREE concentrations, here  
318 considered from Eu to Lu (Fig. 8g), ranging from 490 to 2430 ppm in metatexites, 488  
319 to 1679 ppm in diatexites, and 261 to 2540 ppm in granites. They also show Eu  
320 negative anomalies (metatexites 0.16 – 0.52, diatexites 0.14 – 0.33 and granites 0.02 –  
321 0.16). On the other hand, zircon from the different FCR-LAC rocks show Ce positive  
322 anomalies, suggesting high amounts of Ce in 4<sup>+</sup> valence, which allowed its greater  
323 incorporation (Trail et al., 2012). Most zircon in the metatexites have La below the  
324 detection limit, so to a better estimation of the REE fractionation, Ce/Yb<sub>N</sub> is used. REE

325 fractionation values ( $Ce/Yb_N$ ) in zircon retrieved from the metatexites are the most  
326 fractionated ( $Ce/Yb_N = 0.00001 - 0.05$ ), while zircon from granites and diatexites  
327 clearly yield more elevated values ( $Ce/Yb_N = 0.01 - 0.4$  and  $0.001 - 0.04$ ,  
328 respectively) (Fig. 8g). Overall, Lu/Hf ratios in zircon from the different units are  
329 variable, decreasing from metatexites (0.015), to diatexites (0.006) and granites  
330 (0.003).

331

### 332 5.3. Isotopic geochemistry

333 The radiogenic isotopes presented in this study were recalculated to 320 Ma, which  
334 corresponds to the migmatization age of the autochthonous domain of the CIZ (318 –  
335 325 Ma, Martínez Catalán et al., 2014; Ferreira et al., 2019) (Table 5). While in the  
336 previous section, granites were presented as a single lithological group, in this section  
337 they are presented as syn-tectonic granites and late-tectonic granites (Ferreira et al.,  
338 2019). Likewise, phyllites are presented as phyllite N and phyllite S, depending if they  
339 were collected to the North (N) or South (S) of the anatectic complex.

340

#### 341 5.3.1. Sr, Nd and Hf isotopic compositions

342 All migmatitic and granitic samples, except two, plot in the fourth quadrant  
343 (radiogenic Sr and less radiogenic Nd) of the  $^{87}Sr/^{86}Sr_{320}$  vs.  $\epsilon Nd_{320}$  diagram (Fig. 9a),  
344 implying time-integrated evolution characterized by enrichment of Rb and Nd relatively  
345 to the less incompatible Sr and Sm, respectively.

346 Granites share similar  $\epsilon Nd_{320}$  values (-6.03 to -8.89), but significantly different  
347  $^{87}Sr/^{86}Sr_{320}$  ratios (0.7022 to 0.7126). These compositions are compatible with the  
348 upper continental crust (Chauvel et al., 2014; Hart et al., 1999). Overall, metatexite  
349 compositions are similar to those of granites with  $^{87}Sr/^{86}Sr_{320}$  values ranging between  
350 0.7039 and 0.7124 and  $\epsilon Nd_{320}$  between -5.21 to -8.86.

351 The metasedimentary units, phyllite N ( $^{87}\text{Sr}/^{86}\text{Sr}_{320} = 0.7153$ ;  $\epsilon\text{Nd}_{320} = -10.07$ ) and  
352 phyllite S ( $^{87}\text{Sr}/^{86}\text{Sr}_{320} = 0.7089$ ;  $\epsilon\text{Nd}_{320} = 3.28$ ), have distinct compositions but both plot  
353 in the field defined for the Douro Group (see Teixeira, 2008). The phyllite S is also  
354 strikingly different from the FCR-LAC in terms of its Nd composition. Most granites and  
355 metatexites show similar compositions to that of the syn-tectonic Aguiar da Beira  
356 muscovite-biotite leucogranite (see Costa et al., 2014), while two samples of diatexite  
357 and one sample of metatexite have similarities with the Olo de Sapo gneisses (see  
358 Montero et al., 2017; Fig. 9b).

359 The anatectic units have enriched Hf isotopic compositions that result in  $\epsilon\text{Hf}_{320}$   
360 ranging from -3.42 to -9.25. Except for facies Ily ( $\epsilon\text{Hf}_{320} = -3.99$ ), granites have similar  
361  $\epsilon\text{Hf}_{320}$  compositions (-7.70 to -8.43). The peculiar composition of Ily facies can be  
362 related to the fact that this granite is a 300 Ma late-tectonic granite, whereas the  
363 remaining granites are syn-tectonic, with ages between 313 and 317 Ma (Ferreira et al.  
364 2019). Hf isotopic compositions of the metasedimentary units show distinct  
365 compositions (phyllite N with  $\epsilon\text{Hf}_{320} = -13.68$  and phyllite S with  $\epsilon\text{Hf}_{320} = -7.16$ ). The  
366  $\epsilon\text{Nd}_{320}$  vs.  $\epsilon\text{Hf}_{320}$  diagram shows an overall positive correlation defined by the granites,  
367 metatexites and some diatexites (Fig. 9c). Despite some isotopic variability, two tight  
368 clusters are noticeable in the  $\epsilon\text{Nd}_{320}$  vs.  $\epsilon\text{Hf}_{320}$  diagram, each including granites,  
369 metatexites and type-1 diatexites. Type-2 diatexite reveals a  $\epsilon\text{Nd}_{320}$  composition  
370 between the two clusters.

371 Phyllites, N and S, yield two-stage Nd model ages ( $\text{TDM}_2$ ) of 1.83 Ga and 779 Ma,  
372 respectively. The range of Nd  $\text{TDM}_2$  ages for the metatexites is 1.45 to 1.73 Ga, for the  
373 diatexites is 1.43 to 1.69 Ga, and for the granites is 1.51 to 1.73 Ga. The Hf  $\text{TDM}_2$  ages  
374 are older, displaying the following values: phyllite N (2.16 Ga), phyllite S (1.75 Ga),  
375 metatexites (1.57 to 1.83 Ga), diatexites (1.51 to 1.88 Ga), and granites (1.54 to 1.83  
376 Ga). These Nd and Hf model ages are similar for most of the anatectic complex  
377 lithologies (granites, diatexites, and metatexites). Nd  $\text{TDM}_2$  for the phyllite S sample is

378 different and younger than the remaining analysed lithologies, and also somewhat  
379 different regarding its Hf TDM<sub>2</sub>.

380

### 381 5.3.2. Pb isotopic composition

382 When recalculated to 320 Ma, the Pb isotopic ratios retrieved from phyllites become  
383 clustered onto one group with the remaining lithologies, in opposition to what was  
384 shown for the Sr, Nd, and Hf isotopic systems in section 5.3.1. The <sup>206</sup>Pb/<sup>204</sup>Pb<sub>320</sub> and  
385 <sup>207</sup>Pb/<sup>204</sup>Pb<sub>320</sub> isotopic ratios tend to be almost identical for all lithologies (<sup>206</sup>Pb/<sup>204</sup>Pb<sub>320</sub>  
386 = 17.24 to 18.27 and <sup>207</sup>Pb/<sup>204</sup>Pb<sub>320</sub> = 15.61 to 15.67), with the exception of one  
387 metatexite outlier (Fig. 10a). In comparison, the <sup>208</sup>Pb/<sup>204</sup>Pb<sub>320</sub> isotopic ratio (36.67 to  
388 39.39) is slightly more variable (Fig. 10b).

389

## 390 6. Discussion

391

### 392 6.1. Granite systematics

393 The intimate field relationship between granites, migmatites and metasediments in  
394 the FCR-LAC suggests a genetic link between these lithotypes. This is supported by  
395 the granites plotting close to ternary minimums in the Ab-Or-Q ternary diagram (Fig.  
396 11c; Tuttle and Bowen, 1958; Winter, 2014). This points out to the S-type character of  
397 the studied granites, which is also supported by their peraluminous character (ASI >  
398 1.1) (Fig. 11a) and by their positioning in the A-B granite classification diagram of  
399 Villaseca et al. (1998) (Fig. 11b). Low CaO (0.2 – 1 wt.%), Na<sub>2</sub>O (in general < 3.2  
400 wt.%), and high K<sub>2</sub>O (≈ 5 wt.%) contents, as well as <sup>87</sup>Sr/<sup>86</sup>Sr<sub>320</sub> ratios up to 0.7124 are  
401 also typical of S-type granites (Chappell and White, 1974). <sup>87</sup>Sr/<sup>86</sup>Sr<sub>320</sub> ratios are  
402 somewhat variable (0.7039 – 0.7124), reflecting variability between different granite  
403 facies. This variability, typical of S-type granites (Chappell and White, 2001), can be a



404 consequence of heterogeneous protoliths or reflect isotopic disequilibrium (see below).  
405 The tight clustering of granites and migmatites in the  $\epsilon\text{Nd}_{320}$  vs.  $\epsilon\text{Hf}_{320}$  plot (Fig. 9c) puts  
406 in evidence the cogenetic association amongst these units, which reinforces their  
407 characterisation as S-type granites.

408 The abundant presence of perthites indicates the hypersolvus character of these  
409 granites, compatible with relatively low  $\text{H}_2\text{O}$  pressure, which is also suggested by the  
410 positioning between the ternary minimum defined at 0.1 and 0.2 GPa in the Ab-Or-Q  
411 ternary diagram (Fig. 11c).

412

## 413 6.2. The diatexite-granite link

414 In an anatectic complex, diatexites are regarded as the result of melting processes  
415 which by melt aggregation, migration, and emplacement give rise to granitic intrusive  
416 bodies (Milord et al., 2001). Diatexites in the FCR-LAC are characterized by a  
417 significant compositional variability, which can provide clues on the shared genetic link  
418 with granites.

419 Major and trace element concentrations were used to geochemically separate  
420 distinct types of diatexites: type-1 ( $\text{SiO}_2 = 68 - 74$  wt%) and type-2 ( $\text{SiO}_2 = 74 - 75$   
421 wt%), whereas granites ( $\text{SiO}_2 = 70 - 75$  wt%) are generally less silica-enriched than  
422 type-2 diatexites. Using the Frost et al. (2001) diagram, type-1 diatexites and granites  
423 are alkali-calcic to alkalic, whereas type-2 diatexites range between the alkali-calcic  
424 and calc-alkalic (Fig. 11d). Regarding trace element compositions, type-2 diatexites  
425 also stand out by their lower Rb (145 – 238 ppm) (Fig. 5a) and higher Yb and Lu  
426 concentrations (Fig. 6b), and flat Tb to Yb patterns ( $\text{Tb}/\text{Yb}_N = 0.8 - 1.5$ ) (Fig. 7b), at  
427 odds with what can be observed in the other diatexites and granites (Fig. 7b and 7c)  
428 ( $\text{Tb}/\text{Yb}_N > 2.64$ ). In addition, type-1 diatexites exhibit fractionated HREE ( $\text{Dy}/\text{Yb}_N = 2 -$   
429 5) and Eu negative anomalies (0.2 – 0.7), in contrast with type-2 diatexites, which

430 reveal unfractionated HREE ( $Dy/Yb_N = 0.8 - 1$ ) and Eu anomalies from slightly negative  
431 to strongly positive (0.6 – 4).

432 Comparing both types of diatexites with granites, it is evident the similarity between  
433 granites and type-1 diatexites. Taking into account the differences between these two  
434 types of diatexites, type-2 could correspond to cumulates of quartz and feldspar that  
435 evolved isolated from type-1 diatexites and granites through an earlier fractional  
436 crystallization process (e.g. Brown et al., 2016; Morfin et al., 2014; Sawyer, 1987). The  
437 Hf isotopic signatures of these two types of diatexites are clearly distinct, with  $\epsilon Hf_{320}$  of  
438 the type-2 diatexites being significantly less radiogenic (up to 6 units; Table 5) than the  
439 reported for type-1 diatexites and also for the studied granites. Thus, the Hf isotopic  
440 data reinforces the idea of existing two distinct types of diatexites and of a genetic link  
441 between type-1 diatexites and granites, which is not so obvious when type-2 diatexites  
442 are considered. The role of melting reactions in this variability is potentially significant  
443 and is discussed in the following section.

444 Taking into account the referred isotope similitude with granites (see above), type-1  
445 diatexites are here considered as related to the granites by melt-residuum separation.  
446 This process is evident in the Harker diagrams (Fig. 4), where this type of diatexites are  
447 tendentially more rich in  $FeO^t$ ,  $MgO$ ,  $TiO_2$  and  $Al_2O_3$ , but comparatively depleted in  $K_2O$ ,  
448  $Na_2O$  and  $SiO_2$  than the associated granites.

449

### 450 6.3. Melting reactions

451 The formation of migmatite-granite complexes is the culmination of high-grade  
452 metamorphism during orogenic events. Partial melting is a sequential process where  
453 the production and accumulation of melt gradually increases with time. As prograde  
454 reactions take place, melting of the protolith lead to the formation of metatextitic and  
455 diatextitic migmatites, culminating with a granitic melt at higher melt fractions (Brown,  
456 2013). Melting reactions can take place at different temperatures, pressures and water

457 contents. Typically, pelitic and greywacke protoliths begin to melt at about 650-700 °C  
458 in the presence of an aqueous fluid (fluid-present reactions), producing up to 60 vol.%  
459 of melt (Brown, 2013; Milord et al., 2001; Vielzeuf and Holloway, 1988; Weinberg and  
460 Hasalová, 2015). During the prograde metamorphic path, as temperature increases to  
461 ca. 800-1000 °C, partial melting can occur in the absence of an external fluid, giving  
462 place to a series of fluid-absent reactions by dehydration of hydrous minerals, such as  
463 micas and amphiboles, which progressively increases melt production up to 70 modal  
464 percent melt (Brown, 2013; Bucher and Grapes, 2011; Yakymchuk and Brown, 2014).

465 Field observations combined with major element geochemistry of the FCR-LAC  
466 units show that the most probable protolith of the granitic melts (s.l.) is of sedimentary  
467 origin (see sections 2 and 6.1). Moreover, the major elements geochemistry, in  
468 particular the CaO/Na<sub>2</sub>O ratio, reinforces the idea that FCR-LAC granites are derived  
469 from pelitic melts (CaO/Na<sub>2</sub>O < 0.3; Jung and Pfänder, 2007) (Fig. 12a). The type of  
470 melting reactions can be inferred based on trace-element modelling using Rb, Ba, Sr  
471 (Inger and Harris, 1993) and taking into account the mineral assemblages of the FCR-  
472 LAC rocks (section 3). Yet, the exact reaction cannot be accurately determined given  
473 its high dependence on the chemical and modal composition of the protolith, which in  
474 the present study is thought to be significantly heterogeneous on basis of several  
475 evidences (sections 6.5 and 6.6; Teixeira, 2008). In this way, we can infer that melting  
476 of the pelitic source was mainly controlled by two consecutive reactions leading to the  
477 progressive increase of melt production.

478 It has been shown that fluid-present and fluid-absent reactions produce melts with  
479 distinct characteristics for some key trace-elements, which are constrained by relative  
480 mass proportions of consumed feldspar and mica (Inger & Harris, 1993; Gao et al.,  
481 2017; Dou et al., 2019). The role of these two reaction types on the evolution of the  
482 studied anatectic complex (FCR-LAC) is well evident on diagrams involving elements  
483 compatible with plagioclase or muscovite (Fig. 12b; Fig. 12c).

484 The operation of fluid-present reaction is characterized by the development of  
485 evolution trends with almost constant Rb/Sr ratios but highly variable Sr and Ba  
486 contents (e.g. Gao et al., 2017; Dou et al., 2019), which at FCR-LAC (Fig. 12b and  
487 12c) encompass phyllites and most of the metatexites.

488 Petrographic observations indicate that the metatexite melanosomes are  
489 characterized by a lower plagioclase/muscovite proportion than the phyllites from  
490 where they are considered to have been derived. This suggests that the following fluid-  
491 induced key reaction (Patiño Douce and Harris, 1998; Weinberg and Hasalová, 2015)  
492 occurred during the initial steps of melting:

493

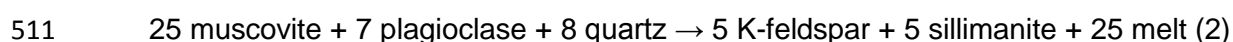


495

496 Increase of Rb/Sr ratios from phyllite, to metatexite, and to diatexite/granite and the  
497 higher Sr and Ba in metatexites in comparison with diatexite/granite (Fig. 12b) seem to  
498 support that early partial melting and melt segregation (phyllites to metatexite  
499 formation) resulted from such congruent fluid-present reaction (1), consuming  
500 plagioclase in larger quantities than muscovite (see also Dou et al., 2019; Gao et al.,  
501 2017; Martini et al., 2019).

502 As temperature increases during prograde metamorphism, fluid-absent melting can  
503 take place by the consumption of hydrous minerals, such as micas. A significant  
504 increase of Rb/Sr ratios accompanied by a decrease of Sr and Ba is observed for  
505 granites and most type-1 diatexites (Fig. 12b and 12c), suggesting dehydration-melting  
506 involving the significant consumption of muscovite and crystallization of peritectic  
507 minerals, such as K-feldspar and sillimanite, in a reaction that could have been similar  
508 to (2) (Patiño Douce and Harris, 1998; see also Gao et al., 2017; Inger and Harris,  
509 1993; Yang et al., 2019):

510



512

513 This reaction also involves the crystallization of peritectic minerals, such as K-  
514 feldspar and sillimanite. This agrees with the absence of primary muscovite in type-1  
515 diatexites and granites, as well as the abundance of K-feldspar and sillimanite.  
516 Reaction (2) also explains the negative Eu anomalies characterizing most of type-1  
517 diatexites and granites ( $\text{Eu}/\text{Eu}^*$  down to 0.2), which can be explained by the  
518 crystallization of peritectic K-feldspar (Fig. 12d; Dou et al., 2019; Gao et al., 2017).

519 After fluid-absent reactions involving the consumption of muscovite, it is commonly  
520 described the dehydration-melting of biotite as a response to temperature increase  
521 along the metamorphic path. There is some petrographic evidence stemming from  
522 migmatites for the breakdown of biotite, such as the occurrence of biotite crystals with  
523 corroded borders. However, reactions involving the melting consumption of biotite are  
524 accompanied by the peritectic production of garnet ( $> 0.4$  GPa) or cordierite ( $< 0.4$   
525 GPa) (Spear et al., 1999; Bento dos Santos et al., 2011). Such minerals were not  
526 detected in the studied samples, although garnet has been reported (Díez Fernández  
527 and Pereira, 2016), and therefore some incipient melting of biotite should have  
528 occurred.

529 We demonstrated before the existence of two geochemically distinct types of  
530 diatexites (see section 6.2) and that type-1 diatexites were derived from fluid absent  
531 reactions (2) associated to the dehydration-melting of muscovite. Type-2 diatexites  
532 exhibit positive Eu anomalies ( $\text{Eu}/\text{Eu}^*$  up to 4), but low CaO contents and Rb/Sr ratios  
533 (Fig. 12b; 12c), compared to the remaining granites and type-1 diatexites. This implies  
534 that peritectic K-feldspar was not formed or only crystallized in small proportions (Dou  
535 et al., 2019; Gao et al., 2017), and, instead, another process must have been  
536 important. Fluid-fluxed melting of muscovite (reaction 1) is a more likely mechanism to  
537 generate melts with compositions similar to type-2 diatexites. The melt productivity of  
538 such reaction is dependent on the amount of fluid present in the rock. The lower  
539 abundance of type-2 diatexites and the lack of granites with compositions compatible

540 with this type of diatexites suggest that the amount of melt generated by that fluid-  
541 present reaction was not enough to promote the melt segregation/extraction needed to  
542 granite generation. This suggests that the amount of fluid in the source rock was small.  
543 Yet, as mentioned in section 6.2, the relative higher silica content in type-2 diatexites  
544 suggests that their composition probably reflect, in addition, the occurrence of some  
545 fractional crystallization.

546 Ti-in-zircon temperatures from young zircon grains within the syn-tectonic granites  
547 range between 783 – 836 °C (Ferreira et al., 2019), implying that the metamorphic  
548 peak reached such temperatures. Additionally, this range of estimated melt  
549 temperatures are consistent with the minimum metamorphic peak temperature  
550 previously calculated from mineral equilibria for the FCR-LAC ( $T = 761 \pm 50$  °C; Pereira  
551 et al., 2017). At this temperature range, up to 60% volume of melt is produced by the  
552 referred partial melting reactions (Vielzeuf and Holloway, 1988), thus allowing  
553 significant melt extraction given that the critical melt percentage for melt migration (20  
554 to 35% volume; e.g. Clemens and Vielzeuf, 1987 and Bento dos Santos et al., 2011a;  
555  $\leq 7\%$ ; Rosenberg and Handy, 2005 and Brown, 2013) is clearly surpassed.

556

#### 557 6.4. The role of accessory minerals

558 Accessory minerals have an important role controlling the composition of trace  
559 elements and isotopic ratios in migmatites and granites (e.g. Ayres and Harris, 1997;  
560 Bea, 1996a, 1996b; Bea et al., 1994; Yakymchuk and Acosta-Vigil, 2019; Zeng et al.,  
561 2005a). Apatite and zircon are the main accessory minerals that occur in the FCR-LAC  
562 rocks and their trace element compositions are described in section 5.2.

563 Results demonstrate that apatite is the main carrier of LREE and Y in the FCR-LAC  
564 (Table 3) and exhibits a flat REE pattern (more evident for LREE; Fig. 8a), typical of  
565 peraluminous rocks (Bea, 1996a). The moderate LREE and  $P_2O_5$  covariation observed

566 in the whole-rock geochemistry (Fig. 13a) confirm the control that apatite has in the  
567 variation of these elements.

568 Conversely, the HREE composition of the anatectic rocks is mainly influenced by  
569 zircon (Table 4). This influence is also confirmed by the covariation between HREE and  
570 Zr/Hf from the whole-rock geochemistry (Fig. 13b-d). Granites show depletion in the  
571 HREE and lower Lu/Hf ratios when compared to metatexites and the country rocks,  
572 which is due to the role of zircon. Indeed, Lu is preferentially retained in restitic zircon  
573 grains during partial melting, resulting in a constant decrease of the Lu concentration  
574 and Lu/Hf ratio in subsequent crystallised zircons.

575 Inherited zircon grains are quite common in all these granites, indicating survival of  
576 significant amounts of zircon, in agreement with partial melting experimental data.  
577 Indeed, considering an average Zr bulk-rock composition of 150 ppm (Table 2), an  
578 estimated maximum melting temperatures of ~840 °C (section 6.3) and an approximate  
579 1:1 proportion of metapelite and greywacke in the protolith, only about 15% of zircon is  
580 expected to dissolve into the melt (Yakymchuk et al., 2017).

581 Apatite grains found in the metasedimentary units are highly enriched in LREE (La  
582 up to 1595 ppm) and Th (up to 97 ppm), which is a typical feature of magmatic apatite  
583 (Henrichs et al., 2018). Therefore, it is plausible that these correspond to detrital  
584 magmatic-derived apatite grains. The recrystallization of apatite during prograde  
585 metamorphism can explain the relative depletion of LREE (La ≤ 405 ppm) in the  
586 metatextitic apatite, typical of metamorphic apatites (Henrichs et al., 2018).

587

#### 588 6.5. Isotope variability

589 The studied rocks form two tight clusters in the  $\epsilon\text{Hf}_{320}$  vs.  $\epsilon\text{Nd}_{320}$  diagram (Fig. 9c) ,  
590 each of them comprising metatexites, diatexites and granites. Sm, Nd, Lu and Hf, the  
591 parent and daughter elements of the Nd and Hf isotopic systems, are all characterized  
592 by high field strengths, which confer them a significant degree of immobility during

593 meteoric and metamorphic processes. Considering this, such clusters clearly suggest  
594 the existence of two somewhat distinct sources for the FCR-LAC anatectic complex, in  
595 agreement with the significant isotopic variability of the DBSG phyllites (Teixeira,  
596 2008). Additionally, the fact that each of the clusters comprise metatexites, diatexites  
597 and granites suggests that during the metatexite-diatexite-granite evolution partial  
598 melting was not accompanied by significant isotopic disequilibrium, like those caused  
599 by the incomplete melting of zircon and the consequent variable release of zircon  
600 unradiogenic  $^{177}\text{Hf}$  and non-zircon Hf during melting (the zircon effect; Tang et al.,  
601 2014). Considering the very long half-life of  $^{176}\text{Lu}$  (37.1 Ga) and  $^{147}\text{Sm}$  (106 Ga) and the  
602 lack of evidence for isotopic disequilibrium, protolith minerals must have had a  
603 comparatively short time period for developing distinct isotopic signatures.  
604 Alternatively, their isotopic compositions may have been homogenised during the high-  
605 grade metamorphism path preceding anatexis itself (Wolf et al., 2019). The occurrence  
606 of restricted isotopic disequilibrium during the first melting stages that led to the  
607 formation of the metatexites should not be discarded. However, given the natural  
608 isotopic variability of the probable source rocks (e.g. Teixeira, 2008) the most likely  
609 explanation for the isotopic variability in the migmatite-granites is source heterogeneity.  
610 Below we will focus on the metatexite-diatexite-granite compositional transitions.

611 It should be noticed that the present-day values ( $\epsilon_0$ ) show a distinct behaviour for the  
612 isotope systems in consideration. Indeed, inside of each of the two groups, enclosing  
613 metatexites/diatexites/granites, a significantly larger variation is observed for the  $\epsilon\text{Hf}_0$   
614 values than for  $\epsilon\text{Nd}_0$  (Fig. 14a). Considering this and the fact that in the  $\epsilon\text{Hf}_{320}$  vs.  
615  $\epsilon\text{Nd}_{320}$  diagram (Fig. 9c) two clusters are observed comprising granites and migmatites,  
616 we propose that a significant fractionation between Hf and Nd isotopic systems  
617 occurred during melting reactions from metatexites to diatexites and finally to granites,  
618 caused by a distinct behaviour of their parent/daughter pairs. As shown before (6.3),  
619 the melting process involved essentially micas and plagioclase, phases with no  
620 capability to significantly fractionate Sm from Nd or Lu from Hf, thus suggesting the role



621 of an accessory phase to explain this different behaviour. Apatite in peraluminous  
622 systems has  $D^{\text{Sm/Nd}} \approx 0.3$  and  $D^{\text{Lu/Hf}} \approx 70$  (Bea et al., 1994), making it a good  
623 candidate. However,  $\epsilon\text{Nd}_0$  of granite-metatexite pairs is similar or only slightly varying,  
624 which would not be expected, and apatite dissolution during melting reactions would  
625 induce an increase of  $^{176}\text{Hf}/^{177}\text{Hf}$  by radiogenic ingrowth of  $^{176}\text{Hf}$  with time, opposite to  
626 what it is observed.. Therefore, decoupling of Lu and Hf cannot be attributed to apatite  
627 dissolution. The observed decoupling of Hf and Nd isotopic systems can be achieved  
628 by the involvement of zircon. Zirconium and Hf are usually considered as two  
629 geochemically coherent elements, which explains the very high  $D_{\text{Hf}}$  characterizing  
630 zircon ( $D_{\text{Hf}} \approx 2420$ ; Wang et al., 2010). The consequent very low Lu/Hf ratios  
631 characterizing zircon will allow, that by incomplete melting of zircon, Lu/Hf in the melt to  
632 be distinct of that of the whole source rock, and that granites would evolve with time to  
633 lower  $^{176}\text{Hf}/^{177}\text{Hf}$  ratios than the less evolved metatexites. Moreover, variable zircon  
634 contribution to melt will contribute to the production of liquid batches characterized by  
635 significantly different Lu/Hf ratios. This would be facilitated by incomplete zircon  
636 dissolution, a process which has been demonstrated to frequently occur in  
637 peraluminous granitic magmas where zircon saturation occurs at relatively low Zr  
638 concentrations ( $\sim 78$  ppm; Harrison and Watson, 1983; Hogan and Sinha, 1991; Tang  
639 et al., 2014) and, also, by the fact that zircon is preferentially located inside biotite  
640 which can protect it from dissolution (Bea, 1996) (see also 6.4). In opposition to the  
641 observed for Lu/Hf ratios, zircon only have a limited capability to fractionate Sm from  
642 Nd ( $D_{\text{Sm/Nd}} \approx 1$ ; Bea et al 1994), thus explaining the contrasting behaviour of Hf and Nd  
643 present day isotope ratios depicted on Fig. 14a).

644 Each of the FCR-LAC metatexite/diatexite/granite groups defined on the  $\epsilon\text{Hf}_{320}$  vs.  
645  $\epsilon\text{Nd}_{320}$  diagram (Fig. 9c) display a significant present-day  $^{87}\text{Sr}/^{86}\text{Sr}$  range at almost  
646 constant  $\epsilon\text{Nd}$  (Fig. 14b). This suggests that Sr isotope signatures evolved under distinct  
647  $^{87}\text{Rb}/^{86}\text{Sr}$ , allowing radiogenic ingrowth of  $^{87}\text{Sr}$  at different rates. At odds with the Nd

648 and Hf systems, which are mainly controlled by the solubility of accessory phases, in  
649 the Rb/Sr system the melt composition is constrained by the major mineral phases  
650 involved in the melting reactions (Tang et al., 2014; Wolf et al., 2019). In fact, an  
651 increase of  $(^{87}\text{Sr}/^{86}\text{Sr})_0$  from metatexites to diatexites and then to granites is observed  
652 (Fig. 14b), which indicates an increase of  $^{87}\text{Rb}/^{86}\text{Sr}$  in the melts progressively  
653 generated at higher temperatures. If we take into account the melting reactions inferred  
654 for the FCR-LAC (as in section 6.3), it is expected firstly a significant contribution of  
655 muscovite to the melt (reactions 1 and 2) and later a minor contribution of biotite, a  
656 mineral usually characterized by significantly higher Rb/Sr than coexisting muscovite  
657 (e.g. Neiva et al., 2002), which together can explain these isotopic ranges and  
658 evolutions.

659 Nonetheless, the range in Sr isotopes is observed even when the initial values  
660  $(^{87}\text{Sr}/^{86}\text{Sr})_{320}$  are considered (Fig. 9a). There are several ways to explain variation on  
661 the initial Sr isotopic composition: isotopic disequilibrium, hydrothermal alteration,  
662 assimilation/fractional crystallization (AFC), and source heterogeneity.

663 Isotopic disequilibrium taking place during the melting stages involving various  
664 minerals characterized by significantly distinct  $^{87}\text{Rb}/^{86}\text{Sr}$  (plagioclase, muscovite,  
665 biotite) is a plausible mechanism, especially in these settings (crustal anatexis).  
666 However, these isotopic signatures had to be developed previously to melting to  
667 account for variable mineral  $^{87}\text{Sr}/^{86}\text{Sr}$  during radiogenic ingrowth (Farina and Stevens,  
668 2011; Zeng et al., 2005b). Yet, at odds with the commonly observed in situations of  
669 isotopic disequilibrium (e.g. Wolf et al., 2019), no systematic isotopic variation is  
670 observed when considered the different lithotypes (metatexites, diatexites, and  
671 granites), but only a significant overlapping of the  $^{87}\text{Sr}/^{86}\text{Sr}_{320}$  characterizing metatexites  
672 and granites. In this perspective, the  $^{87}\text{Sr}/^{86}\text{Sr}_{320}$  variability could likely just be the result  
673 of open-system processes, such as hydrothermal alteration and/or  
674 assimilation/fractional crystallization (AFC). A hydrothermal overprint on these units  
675 would result in a lack of correlation between mobile and high field strength immobile

676 elements, such as LILE and Ti, respectively (e.g. Huang et al., 2019). This is not  
677 supported by our data, which show a clearly linear correlation between K, Sr, Ba vs.  
678  $\text{TiO}_2$  (Fig. 15a-c). Therefore,  $^{87}\text{Sr}/^{86}\text{Sr}_{320}$  variability as a result of post-magmatic  
679 overprint is not plausible.

680 AFC is not a very probable process in constraining the chemical variability of S-type  
681 granitic systems (Clemens and Stevens, 2012; Yakymchuk, 2019). Indeed, as  
682 suggested by Castro (2014), the small difference in density between melt and minerals  
683 precociously crystallised, as well as the high viscosity of high silica melts, lead to  
684 consider as probably small and inefficient the role of fractional crystallization on granitic  
685 magmas. In addition, it has been considered, based on thermodynamics, that granitic  
686 melts have a limited capability of assimilation of crustal rocks, which is more evident for  
687 relatively low-temperature S-type magmas like those in this study (e.g. Glazner, 2007).  
688 In agreement, no positive correlation is observed for the studied rocks between La/Sm  
689 and Th/Sc (Fig. 15d), both ratios being expected to increase during AFC processes  
690 (Huang et al., 2019).

691 Given the lack of evidence for Sr isotopic disequilibrium and for the occurrence of  
692 hydrothermal alteration or AFC processes, the large variability of the  $^{87}\text{Sr}/^{86}\text{Sr}_{320}$  was  
693 most probably caused by source heterogeneity. This is in agreement with geochemical  
694 studies of the DBSG phyllites which have shown a much larger heterogeneity of the Sr  
695 isotope ratios than the observed for  $^{143}\text{Nd}/^{144}\text{Nd}$  (Teixeira, 2008), and in line with  
696 Clemens and Stevens (2012) and Yakymchuk (2019) findings, who concluded that  
697 source heterogeneity is an important factor on isotopic variations in granites in as much  
698 it controls the type of peritectic minerals that will form.

699 On the  $^{207}\text{Pb}/^{204}\text{Pb}$  vs  $^{206}\text{Pb}/^{204}\text{Pb}$  diagram (Fig.14c), the studied rocks define a linear  
700 array close to the evolution line modelled for the upper continental crust by Zartman  
701 and Doe (1981). However, a significantly larger dispersion is observed in the  
702  $^{208}\text{Pb}/^{204}\text{Pb}$  vs.  $^{206}\text{Pb}/^{204}\text{Pb}$  (Fig. 14d and Fig. 10b), suggesting decoupling of  $^{206}\text{Pb}$  and  
703  $^{207}\text{Pb}$  (the final products of the decay chains of  $^{238}\text{U}$  and  $^{235}\text{U}$ , respectively) from  $^{208}\text{Pb}$ ,

704 the final radiogenic product of  $^{232}\text{Th}$ . In granitic systems, the elemental Pb budget is  
705 mainly controlled by K-feldspar, whereas accessory phases are the main repositories  
706 of U and Th (e.g. Bea, 1996), which have the ability to strongly influence the long-term  
707 lead isotope evolution (Hogan and Sinha, 1991; Wolf et al., 2019). Decoupling of  
708 uranogenic lead from thorogenic lead (Fig. 10b) may then reflect the variable modal  
709 proportions of these phases in the source materials and also the variable contribution  
710 to melts of U-rich (e.g.: zircon, apatite and monazite) and Th-rich (e.g.: monazite and  
711 apatite) minerals.

712

#### 713 6.6. Source rocks: some considerations

714 Based on the whole-rock geochemistry (e.g. Harker diagrams; Fig. 4 and 5), we can  
715 distinguish the geochemical evolution due to partial melting and the progressive  
716 residuum-melt separation, with the metapelitic rocks (phyllites) plotting at the origin of  
717 the metatexite-diatexite-granite sequence. This relationship is in clear agreement with  
718 field-based observations (as seen in section 3).

719 Despite the isotopic differences/variability between the phyllite N and phyllite S  
720 (Douro-Beiras Supergroup; Fig. 9b; Teixeira, 2008), and even among the anatectic  
721 complex lithologies, the Sr, Nd, and Hf isotopic compositions show a clear affinity  
722 between migmatites and phyllites. This is more evident for the phyllite N sample with Sr  
723 and Nd signatures within the range of the FCR-LAC units (Fig. 9a and Table 5), than  
724 for the phyllite S sample, where only the Hf initial composition is within analytical  
725 uncertainty of the remaining anatectic units (Fig. 9c and Table 5).  $\text{TDM}_2$  provenance  
726 ages of the metasedimentary protoliths of the anatectic complex broadly agree with the  
727 phyllite units  $\text{TDM}_2$  ages (Table 5), which reinforces the Douro-Beiras Supergroup as  
728 representing the most probable protolith.

729 While we can not go too far beyond our current small dataset to draw further links  
730 between a definitive source to this anatectic complex (either phyllite N or phyllite S), it

731 seems reasonable to propose a stronger affinity with the Douro Group, sampled to the  
732 N. However, we cannot fully discard the involvement of the southern Beiras Group  
733 units (Tassinari et al., 1996) nor a minor contribution from the Olo de Sapo magmatic  
734 rocks (Montero et al., 2017), particularly when the Nd initial isotopic compositions are  
735 taken into account (Fig. 9b). This is particularly the case of the Ily late-tectonic granite,  
736 which exhibits isotopic compositions significantly different from the syn-tectonic  
737 granites (Fig. 9a and 9c; see Ferreira et al., 2019 for granite ages).

738 The contrasting Nd and Hf isotopic results between phyllites sampled from the  
739 northern ( $\epsilon\text{Nd}_{320} = -10.07$ ) and southern ( $\epsilon\text{Nd}_{320} = +3.28$ ) domains are striking, but in  
740 agreement with previously published data for the CIZ metasediments (Villaseca et al.  
741 (2014). This suggests source heterogeneities between both domains, with variable  
742 detrital inputs (more and less evolved crustal sources), which influenced the isotopic  
743 composition of the FCR-LAC units.

744 Further work is needed to clearly demonstrate: a) a systematic isotopic variability  
745 between the metasedimentary units to the North and South of the FCR-LAC; and b) the  
746 affinity of the anatectic complex with the Douro Group. This will further elucidate if the  
747 JPCSZ was a major paleogeographic boundary during the Precambrian/Cambrian, as  
748 proposed by Villaseca et al. (2014) and Iglesias and Ribeiro (1981).

## 749 7. Conclusions

750 This study shows the links between the distinct lithologies of the variscan Figueira  
751 de Castelo Rodrigo – Lumbrales Anatectic Complex of the Central Iberian Zone and  
752 evaluates the occurrence of isotopic disequilibrium during melting reactions from  
753 metatexites to diatexites and granites. The main conclusions are summarized below:

754 1) Based on whole-rock elemental and isotopic geochemistry a genetic link is  
755 definitively established between metatexites, diatexites and granites, which  
756 represent different evolutionary stages of the anatectic complex, culminating on  
757 the genesis of the S-type granites.

758 2) We demonstrate the existence of two geochemically distinct diatexite groups  
759 (type-1 and type-2) generated at different stages of the partial melting process.  
760 We show that type-1 diatexite, with high Rb concentration and fractionated  
761 HREE is geochemically similar to the regional S-type granites, whereas type-2  
762 diatexites are not linked to any of the outcropping granites.

763 3) Based on the mineral assemblage and whole-rock geochemistry we  
764 demonstrated that type-1 diatexites—and granites were produced during  
765 dehydration-melting of muscovite and production of peritectic K-feldspar and  
766 sillimanite plus melt. On the contrary, Rb-poor, unfractionated HREE type-2  
767 diatexites required influx of externally derived fluids during melting, which  
768 consumed more plagioclase than muscovite. The type-2 diatexites represent a  
769 more evolved melt composition than the outcropping granites, being mainly  
770 composed by quartz and feldspar.

771 4) Evidence of incipient biotite melting and Ti-in-zircon thermometry allow to  
772 establish peak temperatures ranging from 783 to 836 °C. These conditions  
773 resulted in the production of large volumes of melt, which explains the profusion  
774 of S-type granites within the CIZ.

775 5) There are no signs of significant isotopic disequilibrium during melting in the  
776 transition from metatexites to diatexites and, finally, to granites. This means that  
777 the source minerals were in isotopic equilibrium at the onset of the melting  
778 process, either reflecting that they were formed shortly before melting, or,  
779 alternatively, that isotopic homogenization during the high-grade metamorphism  
780 preceding the melting process took place. However, the distinct contributions of  
781 mineral phases to the melt produced batches of melt with distinct  
782 parent/daughters' ratios, which resulted in distinct radiogenic ingrowth, and  
783 explain the present-day isotopic heterogeneity found in these units.

784 6) Rb/Sr ratios and the Pb budget of the successively generated lithotypes were  
785 mainly controlled by major minerals such as muscovite, K-feldspar and

786 plagioclase. The accessory phases were the main minerals controlling the  
787 Sm/Nd, Lu/Hf ratios and the U and Th contents.

788 7) Two distinct groups, both comprising metatexites, diatexites and granites, were  
789 identified when comparing the initial isotopic ratios, testifying for the existence of  
790 source heterogeneity. This is consistent with the isotopic variability of  
791 metasediments considered to be the sources of the anatectic complex and may  
792 also explain the differences in the initial Hf and Nd isotopic compositions  
793 between late-tectonic granite-migmatite cluster and the syn-tectonic granite-  
794 migmatite cluster.

795 8) The genetic links with probable protoliths of the anatectic complex were also  
796 explored. We showed that anatectic complex is sourced on the  
797 Neoproterozoic/Lower Cambrian Douro-Beiras Supergroup, with possible minor  
798 contribution of the Ordovician Ollo de Sapo magmatic rocks.

799

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809

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#### 1117 Figure Captions

1118 Fig.1 – (a) Figueira de Castelo Rodrigo-Lumbrals Anatectic Complex in the Iberian  
1119 Variscan Belt setting (adapted from Dias et al., 2016); (b) Geological map of the  
1120 Figueira de Castelo Rodrigo-Lumbrals Anatectic Complex (modified from Silva and  
1121 Ribeiro, 2000).

1122 Fig. 2 – Field observations: (a) phyllites with andalusite porphyroblasts; (b) metatexite  
1123 with bands of peritectic sillimanite; (c) metatexite with boudinated leucosome veins; (d)  
1124 leucosome pocket associated to metatexite; (e) restitic nodule in diatexite; (f) schlieren  
1125 structures in diatexite; (g) nebulitic texture in diatexite; (h) and (i) two-mica granites with  
1126 variable grain size.

1127 Fig. 3 – Petrographic observations: (a) phyllite with foliated texture; (b) pseudomorph of  
1128 andalusite in the phyllite; (c) alternation of leucosome and melanosome in the  
1129 metatexite; (d) fibrolite included in the retrograde secondary muscovite in metatexite;  
1130 (e) Corroded biotite and melt films evidences (red arrows) in diatexite; (f) secondary  
1131 muscovite including fibrolite in diatexite; (g) granite texture and typical mineralogy; (h)  
1132 granite with evidences of deformation.

1133 Fig. 4 – Harker diagrams for major elements for the host rocks (phyllites and  
1134 quartzphyllites) and FCR-LAC lithologies (metatexites, diatexites and granites).

1135 Fig. 5 – Harker diagrams for trace elements for the host rocks (phyllites and  
1136 quartzphyllites) and FCR-LAC lithologies (metatexites, diatexites and granites).

1137 Fig. 6 – (a) Multi-element diagram for phyllites, quartzphyllites and metatexites; (b)  
1138 Multi-element diagram for type-1 diatexites, type-2 diatexites and granites. Chondrite  
1139 normalization values after Palme and O’Neil (2003).

1140 Fig. 7 – (a) REE diagram for phyllites, quartzphyllites and metatexites; (b) REE  
1141 diagram for type-1 diatexites and type-2 diatexites; (c) REE diagram for granites.  
1142 Chondrite normalization values after Palme and O’Neil (2003).

1143 Fig. 8 – (a), (b), (c), (d) and (e) Sr, Y, Th, U and REE (ppm) *versus* SiO<sub>2</sub> WR (Whole-  
1144 Rock; wt. %), respectively; (f) Apatite REE pattern for phyllites, quartzphyllites,  
1145 metatexites, type-1 diatexites, type-2 diatexites and granites; (g) Zircon REE pattern for  
1146 metatexites, type-1 diatexites, type-2 diatexites and granites. Chondrite normalization  
1147 values after Palme and O’Neil (2003).

1148 Fig. 9 – (a)  $^{87}\text{Sr}/^{86}\text{Sr}_{320}$  *versus*  $\epsilon\text{Nd}_{320}$  diagram for the host rocks (phyllites) and FCR-  
1149 LAC lithologies (metatexites, diatexites and granites); (b)  $^{87}\text{Sr}/^{86}\text{Sr}_{320}$  *versus*  $\epsilon\text{Nd}_{320}$   
1150 diagram for the host rocks (phyllites), FCR-LAC lithologies (metatexites, diatexites and  
1151 granites) and comparison with other geological domains of the CIZ: Douro-Beiras  
1152 Supergroup (Douro Group – Teixeira, 2008; Beiras Group – Tassinari et al., 1996;  
1153 South CIZ – Villaseca et al., 2014), Aguiar da Beira granitoids (biotite granodiorite-  
1154 granite and muscovite-biotite leucogranite; Costa et al., 2014), and Ollo de Sapo  
1155 gneisses (Montero et al., 2017). The Aguiar da Beira granitoids were chosen because  
1156 of its proximity to the FCR-LAC and their syn-tectonic character; (c)  $\epsilon\text{Nd}_{320}$  *versus*  
1157  $\epsilon\text{Hf}_{320}$  diagram for the host rocks (phyllites) and FCR-LAC lithologies (metatexites,  
1158 diatexites and granites). The green and the purple ellipses correspond to groups  
1159 defined by very close  $\epsilon\text{Nd}_{320}$  vs.  $\epsilon\text{Hf}_{320}$  compositions.

1160 Fig.10 – (a)  $^{206}\text{Pb}/^{204}\text{Pb}_{320}$  *versus*  $^{207}\text{Pb}/^{204}\text{Pb}_{320}$  for the host rocks (phyllites) and FCR-  
1161 LAC lithologies (metatexites, diatexites and granites); (b)  $^{206}\text{Pb}/^{204}\text{Pb}_{320}$  *versus*

1162  $^{208}\text{Pb}/^{204}\text{Pb}_{320}$  for the host rocks (phyllites) and FCR-LAC lithologies (metatexites,  
1163 diatexites and granites).

1164 Fig. 11 – (a) Aluminium Saturation Index (ASI) classification for granitic and diatexitic  
1165 rocks proposed by (Frost et al., 2001); (b) A-B granite classification diagram of  
1166 Villaseca et al. (1998); (c) Ab-Or-Q ternary diagram (Tuttle and Bowen, 1958; Winter,  
1167 2014) for the FCR-LAC granites; (d)  $\text{Na}_2\text{O}+\text{K}_2\text{O}-\text{CaO}$  (wt. %) versus  $\text{SiO}_2$  (wt. %)  
1168 classification diagram (Frost et al., 2001) showing the ranges between alkalic and  
1169 calcic composition for the FCR-LAC diatexites and granites.

1170 Fig. 12 – (a)  $\text{CaO}/\text{Na}_2\text{O}$  versus  $\text{Al}_2\text{O}_3/\text{TiO}_2$  (wt. %) distinguish between pelite-derived  
1171 melts ( $\text{CaO}/\text{Na}_2\text{O} < 0.3$ ) and melts derived from greywackes or igneous sources (Jung  
1172 and Pfänder, 2007) for FCR-LAC granites and diatexites; (b)  $\text{Rb}/\text{Sr}$  versus  $\text{Sr}$  (ppm)  
1173 and (c)  $\text{Rb}/\text{Sr}$  versus  $\text{Ba}$  (ppm) for the host rocks (phyllites) and FCR-LAC lithologies  
1174 (metatexites, diatexites and granites). The vectors represent the evolution of partial  
1175 melting reactions (Inger and Harris, 1993):  $\text{Mu}(\text{VP})$  – vapour-present muscovite  
1176 melting;  $\text{Mu}(\text{VA})$  – vapour-absent muscovite melting;  $\text{Bi}(\text{VA})$  – vapour-absent biotite  
1177 melting; (d)  $\text{CaO}$  (wt. %) versus  $\text{Eu}/\text{Eu}^*$  for the host rocks (phyllites) and FCR-LAC  
1178 lithologies (metatexites, diatexites and granites).

1179 Fig. 13 – Apatite and zircon role during the partial melting process in the FCR-LAC: (a)  
1180  $\text{P}_2\text{O}_5$  versus LREE (ppm); (b)  $\text{Zr}$  versus  $\text{Hf}$  (ppm); (c)  $\text{Zr}$  versus HREE (ppm); (d)  $\text{Hf}$   
1181 versus HREE (ppm).

1182 Fig. 14 – (a)  $\epsilon\text{Nd}_0$  versus  $\epsilon\text{Hf}_0$  diagram for the host rocks (phyllites) and FCR-LAC  
1183 lithologies (metatexites, diatexites and granites); (b)  $^{87}\text{Sr}/^{86}\text{Sr}_0$  versus  $\epsilon\text{Nd}_0$  diagram for  
1184 the host rocks (phyllites) and FCR-LAC lithologies (metatexites, diatexites and  
1185 granites); (b) and (c) Plumbotectonic model after Zartman and Doe (1981) applied to  
1186 the host rocks (phyllites) and FCR-LAC lithologies (metatexites, diatexites and  
1187 granites): A – mantle; B – orogene; C – upper crust; D – lower crust.

1188 Fig. 15 – Covariation between LILE (K, Sr and Ba) and HFSE (TiO<sub>2</sub>), (a), (b) and (c),  
1189 respectively; (d) AFC assessment – La/Sm *versus* Th/Sc (ppm).

1190

1191 Table 1 – Modal mineralogical proportion for the host rocks (phyllites and  
1192 quartzphyllites) and rocks from the Figueira de Castelo Rodrigo-Lumbrales Anatectic  
1193 Complex (metatexites, diatexites and granites).

1194 Table 2 – Whole-rock chemical composition of phyllites, quartzphyllites (qtzphyllites)  
1195 and FCR-LAC rocks (metatexites, diatexites, and granites). Major elements in weight  
1196 percentage (wt. %) and the trace elements expressed in parts per million (ppm).

1197 Table 3 – Apatite trace element composition (ppm) from the host rocks (phyllites) and  
1198 FCR-LAC rocks (metatexites, diatexites, and granites).

1199 Table 4 – Zircon trace element composition (ppm) from FCR-LAC rocks (metatexites,  
1200 diatexites, and granites).

1201 Table 5 – Whole-rock Sr, Nd, Hf and Pb isotopic data for the host rocks (phyllites) and  
1202 FCR-LAC rocks (metatexites, diatexites, and granites).

Table 1

	Phyllites	Quartzphyllites	Metatexites	Type-1 Diatexites	Type-2 Diatexites	Granites
Quartz	36	48	32	32	40	33
Plagioclase	10	-	14	22	28	23
K-feldspar	-	-	7	12	14	17
Biotite	30	30	22	9	8	12
Muscovite	19	19	21	20	9	12
Sillimanite	-	-	3	4	<1	3
Chlorite	4	2	-	-	-	-
Accessory minerals	<1	<1	<1	<1	<1	<1



Table 2

Lithology	Phyllite N		Phyllite S		Qtzphyllite N	Qtzphyllite S	Metatexite										
	JTJ-2	JTJ-6	JTJ-18	JTJ-19	JTJ-49	JTJ-41b	JTJ-8	JTJ-20	JTJ-22B	JTJ-24B	JTJ-25	JTJ-37	JTJ-48B	JTJ-53	JTJ-54B	JTJ-56A	JTJ-60A
SiO <sub>2</sub>	61.02	54.64	63.82	66.63	68.39	65.71	58.57	66.85	63.74	74.73	60.94	69.23	59.31	64.82	58.51	69.93	71.14
Al <sub>2</sub> O <sub>3</sub>	18.02	23.40	17.68	14.94	14.18	16.15	20.42	16.58	17.78	12.47	18.57	15.08	18.69	15.58	20.42	14.10	15.38
Fe <sub>2</sub> O <sub>3</sub> <sup>1</sup>	6.95	7.34	4.91	5.36	4.55	6.24	8.10	5.51	5.61	3.78	6.52	4.76	7.32	5.90	6.79	4.47	2.50
FeO <sup>1</sup>	6.25	6.60	4.42	4.82	4.09	5.61	7.29	4.96	5.05	3.40	5.87	4.28	6.59	5.31	6.11	4.02	2.25
MnO	0.06	0.07	0.05	0.04	0.06	0.08	0.09	0.06	0.05	0.12	0.04	0.07	0.10	0.10	0.09	0.09	0.08
MgO	2.40	2.06	1.75	1.67	1.65	2.00	2.43	1.74	2.01	1.25	2.24	1.84	2.70	2.16	2.73	1.69	0.73
CaO	1.36	0.30	0.10	0.15	2.36	0.44	0.55	0.79	0.35	0.76	0.51	1.13	0.54	0.23	0.69	0.74	0.60
Na <sub>2</sub> O	1.75	1.14	1.18	1.35	2.52	1.75	1.57	2.30	1.18	2.18	1.40	2.65	1.21	1.45	1.82	2.80	3.04
K <sub>2</sub> O	3.40	4.49	4.34	2.86	3.68	3.03	3.92	3.48	3.84	3.41	4.16	3.27	4.42	2.84	3.67	3.41	4.34
TiO <sub>2</sub>	0.87	0.82	0.76	0.86	0.76	0.89	0.84	0.86	0.70	0.58	0.78	0.64	0.82	0.70	0.74	0.60	0.32
P <sub>2</sub> O <sub>5</sub>	0.14	0.13	0.11	0.07	0.15	0.15	0.13	0.11	0.14	0.17	0.17	0.19	0.15	0.14	0.17	0.16	0.24
LOI	3.65	5.09	4.22	5.02	1.01	3.60	3.72	2.25	3.69	1.23	4.51	1.53	4.43	4.72	4.75	1.63	1.37
Total	99.62	99.45	98.92	98.95	99.31	100.00	100.30	100.50	99.08	100.70	99.84	100.40	99.68	98.63	100.40	99.62	99.74
Sc	15.00	18.00	15.00	14.00	11.00	15.00	18.00	15.00	15.00	8.00	16.00	12.00	17.00	15.00	17.00	13.00	7.00
V	91.00	115.00	123.00	94.00	67.00	98.00	112.00	92.00	111.00	59.00	111.00	79.00	109.00	136.00	124.00	72.00	36.00
Ba	737.00	780.00	764.00	608.00	660.00	608.00	735.00	497.00	615.00	683.00	921.00	664.00	738.00	1038.00	1005.00	361.00	286.00
Sr	209.00	110.00	64.00	49.00	243.00	82.00	118.00	149.00	79.00	156.00	114.00	189.00	111.00	74.00	169.00	107.00	79.00
Y	32.00	29.00	28.00	34.00	28.00	33.00	29.00	22.00	24.00	21.00	26.00	29.00	28.00	33.00	29.00	26.00	14.00
Zr	203.00	133.00	211.00	218.00	386.00	285.00	142.00	188.00	162.00	274.00	142.00	200.00	140.00	155.00	140.00	240.00	114.00
Cr	90.00	110.00	90.00	80.00	60.00	90.00	100.00	80.00	80.00	50.00	90.00	70.00	100.00	100.00	90.00	70.00	40.00
Ni	40.00	40.00	20.00	30.00	20.00	20.00	50.00	30.00	20.00	< 20	30.00	< 20	40.00	50.00	30.00	30.00	< 20
Ga	25.00	32.00	23.00	20.00	19.00	21.00	29.00	22.00	23.00	16.00	25.00	19.00	26.00	23.00	28.00	21.00	22.00
Rb	154.00	184.00	174.00	116.00	154.00	110.00	189.00	167.00	180.00	186.00	170.00	150.00	234.00	170.00	281.00	279.00	260.00
Nb	15.00	14.00	11.00	12.00	13.00	12.00	15.00	13.00	11.00	8.00	12.00	10.00	14.00	11.00	12.00	16.00	19.00
Cs	4.90	9.90	7.60	4.80	8.80	2.10	16.30	12.60	33.30	20.60	9.30	9.40	28.50	13.30	44.10	58.40	42.00
La	51.00	53.90	42.80	46.40	50.90	38.60	61.10	34.70	34.50	43.60	46.40	38.20	45.90	44.70	46.90	40.80	17.70
Ce	92.40	104.00	83.00	93.20	103.00	77.60	114.00	73.00	69.20	91.10	92.10	78.50	90.60	88.30	92.80	82.70	35.00
Pr	11.30	11.90	10.30	10.70	11.40	9.28	13.20	8.22	8.26	9.82	10.90	8.68	10.00	10.30	10.40	9.44	4.03
Nd	40.10	43.30	37.90	40.80	41.60	34.30	46.60	31.00	30.00	35.50	40.90	31.60	37.50	38.40	38.40	34.00	14.00
Sm	7.60	8.40	7.30	8.10	7.60	7.10	8.50	6.20	6.10	6.60	7.90	6.40	7.00	7.40	7.40	6.80	2.90
Eu	1.55	1.69	1.41	1.65	1.67	1.42	1.67	1.23	1.22	1.10	1.50	1.36	1.39	1.47	1.53	1.10	0.54
Gd	6.30	6.80	6.00	7.40	6.30	6.70	6.60	5.20	5.40	5.50	6.60	5.60	6.10	6.80	6.10	5.50	2.50
Tb	1.00	1.00	0.90	1.20	1.00	1.10	1.00	0.80	0.90	0.80	1.00	0.90	1.00	1.00	1.00	0.90	0.50
Dy	6.00	5.90	5.50	6.90	5.70	6.30	6.00	4.50	5.00	4.50	5.60	5.60	5.40	6.00	5.70	5.00	2.70
Ho	1.10	1.10	1.10	1.30	1.10	1.20	1.10	0.80	0.90	0.80	1.00	1.10	1.00	1.20	1.10	1.00	0.50
Er	3.30	3.20	3.10	3.70	3.10	3.70	3.10	2.50	2.80	2.30	2.90	3.40	3.00	3.40	3.10	2.80	1.40
Tm	0.47	0.48	0.46	0.54	0.45	0.53	0.46	0.37	0.41	0.33	0.40	0.51	0.43	0.49	0.43	0.41	0.21
Yb	3.10	3.20	3.10	3.50	3.10	3.50	2.90	2.60	2.90	2.20	2.70	3.40	2.90	3.20	2.90	2.80	1.40
Lu	0.46	0.47	0.46	0.51	0.47	0.55	0.45	0.40	0.45	0.34	0.42	0.53	0.43	0.51	0.44	0.43	0.22
Hf	5.40	3.60	5.20	5.50	9.30	7.00	3.70	4.80	4.20	6.00	3.50	5.00	3.80	3.80	3.80	6.10	2.80
Ta	1.30	1.20	1.10	1.30	1.20	1.00	1.20	1.50	2.10	1.00	0.90	1.10	1.40	0.90	1.00	2.40	4.90
Pb	23.00	21.00	10.00	11.00	22.00	15.00	25.00	20.00	17.00	28.00	21.00	29.00	22.00	22.00	25.00	17.00	21.00
Th	17.10	16.90	12.00	10.20	16.70	13.40	17.20	10.80	10.50	15.30	12.70	11.90	13.20	12.20	13.20	12.80	5.50
U	3.50	6.50	3.60	3.80	3.20	3.80	4.20	5.10	6.60	3.50	4.50	3.90	6.40	43.70	9.40	7.10	3.80
REE	225.68	245.34	203.33	225.90	237.39	191.88	266.68	171.52	168.04	204.49	220.32	185.78	212.65	213.17	218.20	193.68	83.60

LREE	202.40	221.50	181.30	199.20	214.50	166.88	243.40	153.12	148.06	186.62	198.20	163.38	191.00	189.10	195.90	173.74	73.63
HREE	23.28	23.84	22.03	26.70	22.89	25.00	23.28	18.40	19.98	17.87	22.12	22.40	21.65	24.07	22.30	19.94	9.97
LREE/HREE	8.69	9.29	8.23	7.46	9.37	6.68	10.46	8.32	7.41	10.44	8.96	7.29	8.82	7.86	8.78	8.71	7.39
La/Y <sub>N</sub>	11.08	11.34	9.30	8.93	11.06	7.43	14.19	8.99	8.01	13.35	11.57	7.57	10.66	9.41	10.89	9.81	8.51
La/Sm <sub>N</sub>	4.22	4.03	3.69	3.60	4.21	3.42	4.52	3.52	3.56	4.15	3.69	3.75	4.12	3.80	3.98	3.77	3.84
Dy/Y <sub>N</sub>	1.26	1.20	1.15	1.28	1.19	1.17	1.34	1.12	1.12	1.33	1.35	1.07	1.21	1.22	1.28	1.16	1.25
Tb/Y <sub>N</sub>	1.42	1.38	1.28	1.51	1.42	1.38	1.52	1.35	1.37	1.60	1.63	1.16	1.52	1.38	1.52	1.41	1.57
Eu/Eu*	0.68	0.68	0.65	0.65	0.74	0.63	0.68	0.66	0.65	0.56	0.63	0.69	0.65	0.63	0.70	0.55	0.61

Table 2 continued

Lithology	Type-1 Diatexite								Type-2 Diatexite							
	Sample	JTJ-27	JTJ-30	JTJ-31C	JTJ-43	JTJ-48A	JTJ-52	JTJ-55	JPC-48	JTJ-23A	JTJ-24A	JTJ-26	JTJ-28A	JTJ-51	JTJ-56B	JTJ-60B
SiO <sub>2</sub>	72.92	72.29	74.46	68.32	72.75	72.33	73.30	69.83	73.53	74.24	74.82	75.14	75.24	74.98	72.95	74.55
Al <sub>2</sub> O <sub>3</sub>	13.92	14.17	14.07	15.60	15.37	14.94	14.72	15.62	14.35	13.93	13.27	14.96	12.87	12.91	15.25	14.11
Fe <sub>2</sub> O <sub>3</sub>	0.97	1.38	0.90	2.90	1.29	1.52	1.13	2.15	1.12	0.75	0.81	0.18	2.67	1.01	0.98	0.81
FeO <sup>1</sup>	0.87	1.24	0.81	2.61	1.16	1.37	1.02	1.93	1.01	0.67	0.73	0.16	2.40	0.91	0.88	0.73
MnO	0.01	0.01	0.02	0.03	0.03	0.02	0.03	0.03	0.03	0.03	0.01	0.01	0.05	0.03	0.02	0.02
MgO	0.21	0.31	0.25	0.84	0.27	0.41	0.29	0.67	0.18	0.17	0.16	0.06	0.90	0.36	0.49	0.44
CaO	0.59	0.63	0.54	0.95	0.61	0.74	0.88	0.62	0.46	0.59	0.60	0.42	0.65	0.48	0.44	0.21
Na <sub>2</sub> O	3.07	3.12	3.13	2.80	3.39	2.98	3.56	2.91	3.57	3.66	3.15	2.57	2.31	3.16	3.89	2.96
K <sub>2</sub> O	5.31	5.64	5.50	5.71	4.82	5.44	4.18	5.93	4.28	4.97	4.50	6.16	3.77	4.55	3.91	5.27
TiO <sub>2</sub>	0.12	0.20	0.13	0.75	0.19	0.24	0.15	0.40	0.07	0.07	0.11	0.02	0.39	0.12	0.10	0.12
P <sub>2</sub> O <sub>5</sub>	0.36	0.47	0.30	0.44	0.42	0.39	0.25	0.40	0.24	0.25	0.18	0.34	0.14	0.17	0.24	0.20
LOI	1.10	0.93	0.83	1.41	1.05	1.15	1.21	1.23	1.35	0.72	0.98	0.81	1.47	1.50	1.20	1.19
Total	98.58	99.14	100.10	99.74	100.20	100.20	99.68	101.72	99.18	99.38	98.58	100.70	100.50	99.25	99.46	100.61
Sc	4.00	2.00	1.00	3.00	2.00	3.00	3.00	<7	4.00	3.00	5.00	<1	8.00	3.00	3.00	<7
V	<5	<5	<5	29.00	6.00	8.00	<5	13.00	<5	<5	<5	<5	38.00	9.00	9.00	5.00
Ba	196.00	217.00	232.00	640.00	205.00	282.00	468.00	353.00	146.00	203.00	200.00	358.00	582.00	482.00	187.00	220.00
Sr	57.00	56.00	61.00	150.00	50.00	74.00	139.00	89.00	43.00	79.00	63.00	111.00	128.00	115.00	69.00	82.00
Y	12.00	11.00	4.00	11.00	8.00	12.00	9.00	14.00	9.00	8.00	12.00	4.00	24.00	9.00	7.00	10.00
Zr	47.00	79.00	45.00	415.00	74.00	100.00	48.00	163.00	28.00	42.00	35.00	25.00	159.00	42.00	39.00	37.00
Cr	<20	<20	<20	<20	<20	<20	<20	19.00	<20	<20	<20	<20	40.00	<20	<20	6.00
Ni	<20	<20	<20	<20	<20	<20	<20	7.00	<20	<20	<20	<20	<20	<20	<20	<7
Ga	18.00	21.00	19.00	26.00	24.00	20.00	17.00	23.00	16.00	12.00	16.00	17.00	16.00	12.00	15.00	14.00
Rb	239.00	303.00	275.00	331.00	315.00	282.00	143.00	290.00	238.00	170.00	183.00	234.00	165.00	154.00	145.00	205.00
Nb	6.00	7.00	4.00	6.00	10.00	7.00	6.00	10.37	9.00	4.00	3.00	<1	7.00	5.00	5.00	4.98
Cs	10.20	8.60	6.60	7.80	20.30	17.10	14.00	-	34.80	9.70	5.00	6.00	10.10	8.60	6.00	-
La	7.20	15.60	5.90	145.00	14.70	20.40	21.00	36.96	3.50	3.60	4.60	1.10	36.60	7.70	6.80	3.79
Ce	15.40	35.10	13.40	368.00	32.20	44.70	41.70	79.12	7.80	7.20	9.60	2.20	74.20	15.20	13.10	7.77
Pr	1.85	4.23	1.65	46.50	3.78	5.19	4.80	9.79	0.88	0.79	1.11	0.28	8.21	1.64	1.49	0.95
Nd	6.70	16.00	6.10	175.00	14.40	19.50	17.10	36.54	3.40	2.90	4.10	1.10	29.20	6.60	5.20	3.46
Sm	1.80	4.10	1.80	25.40	3.90	5.10	4.10	8.99	0.80	0.70	1.10	0.40	5.90	1.30	1.00	0.98
Eu	0.41	0.45	0.39	1.06	0.33	0.55	0.79	0.61	0.25	0.40	0.44	0.55	1.08	0.56	0.34	0.43
Gd	1.90	3.60	1.80	9.50	3.50	4.70	3.30	6.86	1.00	0.80	1.30	0.50	4.80	1.30	0.90	1.12
Tb	0.40	0.60	0.30	0.80	0.50	0.70	0.50	0.85	0.30	0.20	0.30	0.10	0.80	0.20	0.20	0.24
Dy	2.20	2.90	1.10	3.10	2.10	3.10	2.10	3.52	1.70	1.30	2.00	0.60	4.70	1.40	1.10	1.70

Ho	0.40	0.40	0.20	0.40	0.20	0.40	0.30	0.49	0.30	0.30	0.40	0.10	0.90	0.20	0.20	0.35
Er	1.10	0.70	0.30	0.90	0.50	0.80	0.60	1.10	0.90	0.80	1.20	0.30	2.50	0.70	0.70	1.07
Tm	0.14	0.08	0.05	0.12	0.06	0.11	0.08	0.15	0.14	0.14	0.20	0.05	0.37	0.10	0.11	0.18
Yb	0.90	0.40	0.20	0.70	0.40	0.60	0.50	1.00	1.00	1.10	1.40	0.30	2.30	0.80	0.90	1.27
Lu	0.12	0.01	0.01	0.09	0.05	0.10	0.06	0.14	0.15	0.17	0.22	0.04	0.36	0.13	0.13	0.20
Hf	1.40	2.10	1.30	10.00	2.10	2.80	1.50	2.32	0.80	1.20	0.90	0.90	4.10	1.00	1.30	<0.06
Ta	0.70	0.80	0.60	0.50	1.90	0.90	1.10	1.03	1.90	1.30	0.30	< 0.1	0.90	1.00	1.40	0.37
Pb	32.00	33.00	33.00	41.00	27.00	36.00	36.00	39.00	36.00	41.00	31.00	43.00	28.00	37.00	28.00	38.00
Th	2.50	7.50	2.90	133.00	6.90	8.10	6.90	20.62	0.90	0.90	1.30	< 0.1	12.30	2.30	1.80	1.54
U	5.50	9.30	6.10	10.30	10.00	7.80	5.90	13.48	6.80	12.80	3.30	1.50	5.40	3.00	7.10	3.38
REE	40.52	84.17	33.20	776.57	76.62	105.95	96.93	186.13	22.12	20.40	27.97	7.62	171.92	37.83	32.17	23.52
LREE	32.95	75.03	28.85	759.90	68.98	94.89	88.70	171.40	16.38	15.19	20.51	5.08	154.11	32.44	27.59	16.96
HREE	7.57	9.14	4.35	16.67	7.64	11.06	8.23	14.73	5.74	5.21	7.46	2.54	17.81	5.39	4.58	6.57
LREE/HREE	4.35	8.21	6.63	45.58	9.03	8.58	10.78	11.64	2.85	2.92	2.75	2.00	8.65	6.02	6.02	2.58
La/Yb <sub>N</sub>	5.39	26.27	19.87	139.50	24.75	22.90	28.29	24.80	2.36	2.20	2.21	2.47	10.72	6.48	5.09	2.01
La/Sm <sub>N</sub>	2.51	2.39	2.06	3.59	2.37	2.51	3.22	2.58	2.75	3.23	2.63	1.73	3.90	3.72	4.27	2.43
Dy/Yb <sub>N</sub>	1.59	4.71	3.57	2.88	3.41	3.36	2.73	2.28	1.10	0.77	0.93	1.30	1.33	1.14	0.79	0.87
Tb/Yb <sub>N</sub>	1.96	6.60	6.60	5.03	5.50	5.13	4.40	3.72	1.32	0.80	0.94	1.47	1.53	1.10	0.98	0.83
Eu/Eu*	0.68	0.36	0.66	0.21	0.27	0.34	0.66	0.24	0.85	1.63	1.12	3.76	0.62	1.32	1.10	1.26

Table 2 continued

Lithology	Granite																
	Sample	JTJ-31A	JTJ-32	JTJ-34	JTJ-40	JPC-49	JPC-56	JPC-57	JTJ-9	JTJ-10	JTJ-44	JTJ-45	JTJ-47	JTJ-50	JTJ-62	JPC-51	JPC-52
SiO <sub>2</sub>	73.04	71.18	70.97	74.07	71.66	72.13	71.99	73.37	71.55	71.54	73.60	70.78	72.76	73.31	71.53	71.37	
Al <sub>2</sub> O <sub>3</sub>	15.07	14.88	15.14	14.33	14.95	14.88	15.02	14.56	14.70	15.20	14.40	14.89	14.50	15.40	14.92	15.19	
Fe <sub>2</sub> O <sub>3</sub> <sup>I</sup>	1.04	1.91	2.36	0.51	1.94	1.56	1.50	1.28	1.45	1.56	1.36	1.90	1.32	0.92	1.48	1.41	
FeO <sup>I</sup>	0.94	1.72	2.12	0.46	1.75	1.40	1.35	1.15	1.30	1.40	1.22	1.71	1.19	0.83	1.33	1.27	
MnO	0.02	0.03	0.02	0.01	0.02	0.02	0.02	0.02	0.03	0.02	0.02	0.03	0.03	0.04	0.02	0.02	
MgO	0.26	0.46	0.57	0.09	0.60	0.65	0.65	0.25	0.29	0.43	0.32	0.48	0.23	0.10	0.71	0.60	
CaO	0.55	0.71	0.88	0.34	0.45	0.36	0.36	0.53	0.62	0.60	0.55	0.71	0.57	0.44	0.34	0.33	
Na <sub>2</sub> O	2.90	2.51	2.61	1.99	2.67	3.14	3.03	2.98	3.30	3.04	2.79	2.83	3.15	3.92	3.87	3.04	
K <sub>2</sub> O	5.24	5.43	5.98	7.57	5.61	5.38	5.54	4.99	5.02	5.34	5.28	5.56	5.01	4.05	5.47	6.09	
TiO <sub>2</sub>	0.15	0.29	0.54	0.06	0.32	0.27	0.25	0.19	0.19	0.27	0.21	0.33	0.16	0.08	0.22	0.27	
P <sub>2</sub> O <sub>5</sub>	0.38	0.48	0.41	0.55	0.33	0.37	0.39	0.36	0.40	0.44	0.41	0.35	0.35	0.45	0.38	0.33	
LOI	0.94	1.28	1.20	0.90	1.34	1.09	1.13	1.61	0.96	1.25	1.30	1.11	0.96	1.23	0.98	1.19	
Total	99.59	99.15	100.70	100.40	101.64	101.25	101.23	100.10	98.52	99.67	100.20	98.96	99.01	99.95	101.25	101.11	
Sc	1.00	3.00	2.00	< 1	< 7	< 7	< 7	2.00	2.00	2.00	3.00	2.00	2.00	2.00	< 7	< 7	
V	8.00	10.00	17.00	< 5	11.00	7.00	8.00	7.00	5.00	11.00	7.00	13.00	< 5	< 5	7.00	8.00	
Ba	211.00	279.00	357.00	143.00	258.00	235.00	217.00	243.00	225.00	262.00	206.00	263.00	139.00	46.00	213.00	261.00	
Sr	60.00	68.00	84.00	54.00	69.00	67.00	62.00	57.00	57.00	64.00	54.00	80.00	48.00	27.00	62.00	79.00	
Y	5.00	17.00	10.00	9.00	12.00	10.00	12.00	7.00	7.00	10.00	7.00	7.00	6.00	6.00	10.00	9.00	
Zr	57.00	137.00	287.00	23.00	129.00	108.00	104.00	64.00	70.00	106.00	84.00	117.00	64.00	40.00	96.00	114.00	
Cr	< 20	< 20	< 20	< 20	8.00	8.00	7.00	< 20	< 20	< 20	< 20	< 20	< 20	< 20	5.00	5.00	
Ni	< 20	< 20	< 20	< 20	< 7	< 7	< 7	< 20	< 20	< 20	< 20	< 20	< 20	< 20	< 7	< 7	
Ga	23.00	24.00	27.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	22.00	27.00	23.00	26.00	20.00	23.00	
Rb	325.00	316.00	376.00	401.00	284.00	343.00	311.00	311.00	342.00	330.00	338.00	416.00	419.00	521.00	339.00	371.00	

Nb	5.00	9.00	7.00	3.00	10.54	11.26	11.68	9.00	9.00	7.00	9.00	9.00	14.00	15.00	10.08	8.87
Cs	20.30	18.40	11.50	13.60	-	-	-	13.30	27.00	15.20	18.90	17.80	26.60	37.50	-	-
La	10.80	31.80	97.30	2.90	37.13	24.76	22.92	15.50	15.90	21.40	19.80	36.00	15.00	8.00	21.82	31.72
Ce	25.00	75.50	248.00	7.80	85.76	54.74	52.13	34.00	34.60	50.80	47.50	84.10	34.10	17.90	50.56	71.00
Pr	3.08	9.37	31.50	1.05	11.01	6.94	6.70	4.01	3.98	6.28	5.85	10.30	4.08	2.13	6.57	9.03
Nd	12.40	35.40	119.00	4.30	41.49	26.35	25.79	14.80	15.20	24.10	22.40	38.50	14.50	7.80	25.37	34.11
Sm	3.10	8.10	20.50	1.50	8.33	6.07	6.00	3.70	3.80	5.80	4.50	8.50	3.60	1.90	5.77	7.68
Eu	0.41	0.54	0.78	0.33	0.42	0.42	0.40	0.34	0.36	0.51	0.38	0.62	0.33	0.13	0.37	0.54
Gd	2.30	6.20	9.30	1.50	5.12	4.50	4.59	3.40	3.50	4.60	3.20	5.40	2.80	1.60	3.96	5.16
Tb	0.30	0.90	0.80	0.30	0.63	0.54	0.63	0.50	0.50	0.60	0.40	0.60	0.40	0.30	0.51	0.58
Dy	1.20	4.00	2.90	1.70	2.75	2.18	2.82	1.90	2.00	2.40	1.80	2.00	1.80	1.30	2.29	2.18
Ho	0.20	0.60	0.30	0.30	0.39	0.29	0.38	0.20	0.20	0.30	0.20	0.20	0.20	0.20	0.32	0.29
Er	0.40	1.30	0.70	0.70	0.84	0.58	0.78	0.40	0.40	0.60	0.60	0.60	0.50	0.50	0.70	0.56
Tm	0.05	0.15	0.09	0.08	0.09	0.07	0.10	0.06	0.05	0.07	0.07	0.06	0.06	0.06	0.08	0.07
Yb	0.30	0.80	0.50	0.50	0.60	0.38	0.55	0.30	0.30	0.40	0.40	0.30	0.30	0.40	0.47	0.43
Lu	0.01	0.10	0.07	0.06	0.08	0.05	0.06	0.04	0.04	0.06	0.06	0.04	0.04	0.06	0.06	0.06
Hf	1.50	3.70	7.00	1.00	1.55	<0.06	1.50	1.70	1.90	2.80	2.20	3.10	1.80	1.50	1.31	3.02
Ta	0.90	1.10	0.80	0.40	0.58	1.14	1.14	1.80	1.90	1.20	1.50	1.00	3.30	4.90	1.20	0.86
Pb	28.00	36.00	39.00	43.00	38.00	35.00	35.00	27.00	27.00	31.00	29.00	36.00	32.00	16.00	36.00	43.00
Th	6.80	22.20	107.00	1.20	39.25	22.16	20.21	6.90	7.20	13.90	15.60	28.50	9.10	5.10	20.86	33.69
U	6.70	12.60	10.70	3.50	16.30	11.93	14.03	5.10	9.40	8.70	11.10	15.00	13.70	4.50	10.69	5.84
REE	59.55	174.76	531.74	23.02	194.63	127.86	123.85	79.15	80.83	117.92	107.16	187.22	77.71	42.28	118.85	163.41
LREE	54.38	160.17	516.30	17.55	183.71	118.86	113.54	72.01	73.48	108.38	100.05	177.40	71.28	37.73	110.09	153.54
HREE	5.17	14.59	15.44	5.47	10.92	9.00	10.31	7.14	7.35	9.54	7.11	9.82	6.43	4.55	8.76	9.88
LREE/HREE	10.52	10.98	33.44	3.21	16.83	13.20	11.01	10.09	10.00	11.36	14.07	18.07	11.09	8.29	12.57	15.55
La/Yb <sub>N</sub>	24.24	26.77	131.06	3.91	41.48	43.79	27.82	34.80	35.69	36.03	33.34	80.82	33.67	13.47	31.35	49.25
La/Sm <sub>N</sub>	2.19	2.47	2.98	1.22	2.80	2.56	2.40	2.63	2.63	2.32	2.77	2.66	2.62	2.65	2.38	2.59
Dy/Yb <sub>N</sub>	2.60	3.25	3.77	2.21	2.96	3.72	3.30	4.11	4.33	3.90	2.92	4.33	3.90	2.11	3.18	3.27
Tb/Yb <sub>N</sub>	4.40	4.95	7.04	2.64	4.57	6.24	5.00	7.33	7.33	6.60	4.40	8.80	5.87	3.30	4.82	5.83
Eu/Eu*	0.47	0.23	0.17	0.67	0.20	0.24	0.24	0.29	0.30	0.30	0.31	0.28	0.32	0.23	0.24	0.26

Table 2 continued

Lithology	Granite														
	Sample	JPC-53	JTJ-35	JPC-50	JTJ-11	JTJ-12	JPC-55	JTJ-59	JTJ-13	JPC-19	JPC-24	JPC-59	JPC-54	JTJ-36	JTJ-58
SiO <sub>2</sub>	71.62	70.63	71.41	71.13	71.39	70.32	71.96	71.78	70.64	72.31	75.24	73.33	72.64	73.41	
Al <sub>2</sub> O <sub>3</sub>	15.20	15.10	15.05	14.71	15.01	15.51	14.50	14.60	15.15	14.64	13.95	14.72	14.91	14.81	
Fe <sub>2</sub> O <sub>3</sub> <sup>I</sup>	1.72	2.14	1.51	1.64	1.84	2.15	1.66	1.63	1.73	1.89	0.70	1.19	1.34	0.82	
FeO <sup>I</sup>	1.55	1.93	1.36	1.48	1.66	1.93	1.49	1.47	1.56	1.70	0.63	1.07	1.21	0.74	
MnO	0.02	0.02	0.02	0.02	0.02	0.03	0.04	0.02	0.02	0.03	0.02	0.02	0.02	0.03	
MgO	0.70	0.59	0.55	0.43	0.41	0.75	0.54	0.36	0.63	0.64	0.36	0.48	0.31	0.19	
CaO	0.40	0.72	0.26	0.73	0.61	0.58	1.01	0.58	0.46	0.37	0.09	0.24	0.63	0.55	
Na <sub>2</sub> O	3.02	2.57	2.99	3.06	2.72	3.16	3.54	2.66	2.63	2.80	3.48	2.98	3.25	3.27	
K <sub>2</sub> O	5.53	5.37	5.93	5.43	5.39	5.32	4.86	5.71	6.91	5.39	4.33	5.20	4.97	4.59	
TiO <sub>2</sub>	0.31	0.44	0.22	0.28	0.28	0.42	0.28	0.28	0.40	0.31	0.10	0.19	0.19	0.10	
P <sub>2</sub> O <sub>5</sub>	0.37	0.48	0.47	0.37	0.40	0.39	0.29	0.49	0.31	0.36	0.30	0.30	0.44	0.35	
LOI	0.95	1.51	1.44	0.94	1.06	1.21	1.02	1.21	0.84	1.01	1.32	1.21	1.11	1.67	

Total	101.39	99.58	101.21	98.71	99.14	101.77	99.70	99.31	101.28	101.45	100.52	100.93	99.82	99.78
Sc	<7	3.00	<7	2.00	2.00	<7	3.00	2.00	<7	<7	<7	<7	2.00	3.00
V	8.00	19.00	<5	11.00	9.00	18.00	17.00	10.00	8.00	9.00	<5	5.00	7.00	< 5
Ba	240.00	376.00	183.00	283.00	235.00	287.00	491.00	268.00	407.00	172.00	134.00	146.00	274.00	201.00
Sr	76.00	93.00	47.00	80.00	63.00	74.00	142.00	76.00	106.00	49.00	42.00	46.00	74.00	61.00
Y	10.00	10.00	14.00	9.00	11.00	10.00	8.00	8.00	9.00	11.00	5.00	6.00	9.00	7.00
Zr	123.00	211.00	86.00	112.00	116.00	198.00	130.00	113.00	193.00	130.00	35.00	72.00	74.00	46.00
Cr	9.00	< 20	6.00	< 20	< 20	11.00	< 20	< 20	6.00	7.00	<5	5.00	< 20	< 20
Ni	<7	< 20	<7	< 20	< 20	<7	< 20	< 20	<7	<7	<7	<7	< 20	< 20
Ga	24.00	25.00	23.00	22.00	22.00	25.00	23.00	25.00	23.00	24.00	20.00	22.00	23.00	18.00
Rb	371.00	358.00	403.00	334.00	357.00	369.00	276.00	381.00	305.00	378.00	277.00	293.00	288.00	281.00
Nb	11.61	7.00	14.99	8.00	10.00	9.94	10.00	8.00	4.32	15.97	12.91	7.92	10.00	9.00
Cs	-	10.60	-	14.80	18.30	-	19.10	13.00	-	-	-	-	17.00	18.90
La	33.65	65.40	16.95	26.60	30.40	64.65	38.80	31.20	61.40	37.77	5.03	14.98	12.40	8.50
Ce	75.62	162.00	39.29	60.00	72.70	151.70	77.60	72.00	135.11	87.83	11.04	33.71	26.90	17.70
Pr	9.74	20.40	5.00	7.08	8.82	19.91	8.80	8.88	16.41	11.35	1.36	4.24	3.27	2.09
Nd	36.91	75.60	18.49	26.50	33.50	75.41	30.80	33.80	61.23	43.58	5.09	15.65	12.40	7.50
Sm	8.17	12.90	3.94	6.30	7.00	12.56	5.50	7.40	13.18	9.40	1.22	3.56	3.30	2.20
Eu	0.50	0.74	0.25	0.57	0.45	0.59	0.75	0.56	0.79	0.46	0.11	0.23	0.34	0.51
Gd	5.30	6.40	2.91	4.50	5.10	6.18	3.70	4.80	7.47	5.84	1.20	2.62	3.00	2.10
Tb	0.61	0.70	0.50	0.60	0.70	0.68	0.40	0.60	0.70	0.72	0.19	0.34	0.50	0.30
Dy	2.43	2.80	2.66	2.20	2.80	2.60	1.90	2.10	2.39	2.92	0.95	1.46	2.30	1.80
Ho	0.30	0.40	0.41	0.30	0.40	0.38	0.30	0.30	0.30	0.43	0.15	0.18	0.30	0.30
Er	0.60	0.80	0.91	0.60	0.80	0.82	0.60	0.60	0.69	0.95	0.32	0.40	0.60	0.60
Tm	0.07	0.11	0.11	0.07	0.10	0.10	0.09	0.07	0.06	0.11	0.04	0.04	0.08	0.08
Yb	0.39	0.60	0.64	0.40	0.50	0.62	0.60	0.30	0.46	0.66	0.27	0.26	0.50	0.50
Lu	0.05	0.09	0.08	0.06	0.07	0.08	0.09	0.05	0.05	0.07	0.04	0.03	0.06	0.07
Hf	2.21	5.10	0.48	2.80	3.10	5.09	3.40	2.90	2.18	2.09	1.06	<0.06	2.10	1.50
Ta	1.05	0.80	1.61	1.10	1.60	1.27	1.40	0.90	0.52	1.23	2.09	1.07	2.10	2.50
Pb	39.00	34.00	34.00	36.00	33.00	36.00	34.00	37.00	53.00	34.00	21.00	33.00	26.00	42.00
Th	32.41	57.10	21.08	14.00	22.30	66.13	17.00	24.60	38.52	36.34	2.48	14.02	4.90	3.40
U	19.28	11.60	12.44	9.90	10.70	13.69	10.30	10.60	10.11	13.20	3.67	13.15	5.70	5.80
REE	174.34	348.94	92.13	135.78	163.34	336.28	169.93	162.66	300.24	202.10	27.00	77.70	65.95	44.25
LREE	164.08	336.30	83.67	126.48	152.42	324.23	161.50	153.28	287.33	189.93	23.74	72.14	58.27	37.99
HREE	10.26	12.64	8.46	9.30	10.92	12.05	8.43	9.38	12.91	12.17	3.27	5.56	7.68	6.26
LREE/HREE	15.99	26.61	9.89	13.60	13.96	26.90	19.16	16.34	22.27	15.61	7.27	12.98	7.59	6.07
La/Yb <sub>N</sub>	58.29	73.41	17.92	44.79	40.95	69.91	43.55	70.04	90.16	38.33	12.66	38.24	16.70	11.45
La/Sm <sub>N</sub>	2.59	3.19	2.70	2.65	2.73	3.23	4.43	2.65	2.93	2.53	2.60	2.64	2.36	2.43
Dy/Yb <sub>N</sub>	4.06	3.03	2.71	3.57	3.64	2.71	2.06	4.55	3.39	2.86	2.31	3.58	2.99	2.34
Tb/Yb <sub>N</sub>	6.94	5.13	3.45	6.60	6.16	4.82	2.93	8.80	6.72	4.78	5.69	3.12	4.40	2.64
Eu/Eu*	0.23	0.25	0.23	0.33	0.23	0.20	0.51	0.29	0.24	0.19	0.28	0.23	0.33	0.73

Table 3

Lithology	Phyllite			Metatexite					Type-1 Diatexite				Type-2 Diatexite	Granite X	Granite IX	Granite VIII	Granite V	Granite III	Granite II
Sample	M1_2_1	M1_2_2	M1_2_3	M1_20_1	M2_25_1	M3_8_1	M4_60A_1	M5_53_1	M2_43_1	M2_43_2	M3_55_1	M3_55_2	M1_24A_1	M4_34_1	M3_10_1	M2_35_1	M1_12_1	M5_13_1	M4_36_1
Sr	317	1471	333	167	243	487	113	183	112	115	144	131	86	78	99	96	74	76	133
Y	854	480	697	1973	1915	837	1618	1457	847	852	1569	1634	2026	908	542	1026	1546	1083	212
Th	27	97	18	1	1	50	1	0	15	21	1	0	1	7	5	5	19	5	4
U	8	34	18	223	32	17	91	46	33	47	135	123	185	49	51	62	74	75	81
La	386	1585	897	162	167	405	159	126	496	538	192	169	214	391	330	344	380	340	276
Ce	1124	2019	1672	547	559	921	503	425	1848	1924	654	557	574	1404	879	1230	1289	1129	561
Pr	175	174	179	91	97	107	83	74	329	336	114	96	79	241	120	214	215	181	66
Nd	931	630	747	513	565	455	457	435	1760	1808	643	533	348	1282	514	1115	1114	899	259
Sm	253	94	143	216	245	120	190	195	491	502	321	261	142	432	168	377	414	363	67
Eu	21	21	19	17	26	29	16	17	10	10	14	14	11	10	17	13	8	11	22
Gd	252	98	158	301	350	145	255	278	315	321	444	361	194	334	160	318	392	345	59
Tb	33	16	22	60	67	27	50	52	39	39	82	69	48	45	28	47	66	54	10
Dy	169	100	130	386	410	166	316	320	181	184	404	378	341	205	124	226	327	255	46
Ho	29	18	25	77	78	31	61	60	29	30	55	58	67	31	15	35	49	36	6
Er	67	42	60	199	190	78	155	146	68	68	101	130	192	64	31	75	104	73	12
Tm	8	5	8	28	25	11	22	19	9	9	11	17	32	8	4	9	13	8	2
Yb	51	33	46	183	148	73	145	120	52	53	64	111	231	41	24	53	70	46	10
Lu	7	4	7	25	19	9	19	16	7	7	8	14	31	5	3	6	8	5	1
LREE	2870	4503	3638	1529	1632	2008	1391	1255	4924	5108	1923	1615	1357	3750	2011	3279	3413	2911	1229
HREE	638	338	474	1275	1311	569	1039	1028	709	720	1183	1153	1147	742	407	783	1036	833	168
REE	3508	4841	4112	2804	2943	2577	2431	2283	5633	5829	3106	2768	2504	4493	2418	4062	4448	3744	1397
Eu/Eu*	0.25	0.67	0.39	0.20	0.27	0.67	0.22	0.22	0.07	0.08	0.12	0.14	0.19	0.08	0.33	0.11	0.06	0.09	1.06
La/Yb <sub>N</sub>	5.1	32.6	13.0	0.6	0.8	3.7	0.7	0.7	6.4	6.8	2.0	1.0	0.6	6.5	9.1	4.3	3.7	5.0	19.1
La/Sm <sub>N</sub>	1.0	10.6	3.9	0.5	0.4	2.1	0.5	0.4	0.6	0.7	0.4	0.4	0.9	0.6	1.2	0.6	0.6	0.6	2.6
Dy/Yb <sub>N</sub>	2.1	2.0	1.8	1.4	1.8	1.5	1.4	1.7	2.3	2.2	4.1	2.2	1.0	3.3	3.3	2.8	3.0	3.6	3.1

Table 4

Lithology	Metatexite											Type-1 Diatexite		Type-2 Diatexite	Granite X										Granite IX	Granite V
Sample	M1T_20_11c-1	M1T_20_25c-1	M1T_20_46c-1	M1T_20_57c-1	M1T_20_78c-1	M6T_25_1c-1	M6T_25_2c-1	M6T_25_3c-1	M6T_25_4c-1	M6T_25_5c-1	M6T_25_7c-1	M6T_55_10c-1	M6T_55_12c-1	M1T_24A_17c-1	M6T_34_10c-1	M6T_34_15c-1	M6T_34_23c-1	M6T_34_24c-1	M6T_34_26c-1	M6T_34_27c-1	M6T_34_3c-1	M6T_45_12_1c-1	M6T_12_1c-1			
Hf	10140	11080	8180	12530	10000	14300	13290	13260	13300	14410	14360	12750	11060	11810	12850	11660	11990	12070	12430	12670	12650	10280	11950			
U	1152	341	97	2171	851	1310	1137	1261	1378	1213	1239	475	1514	321	734	1258	1182	1603	1480	1000	190	898	2770			
Th	419	288	57	520	253	4	3	6	4	4	5	28	117	37	84	314	284	398	7400	980	68	214	16300			
Th/U	0.36	0.85	0.58	0.24	0.30	0.003	0.003	0.005	0.003	0.003	0.004	0.06	0.08	0.12	0.11	0.25	0.24	0.25	5.00	0.98	0.36	0.24	5.88			
Y	-	-	677	-	120	-	-	-	-	-	2074	-	-	610	469	855	797	1199	985	592	799	-	-			
La	56	0.2	<LOD	14	3	<LOD	<LOD	<LOD	<LOD	<LOD	1	0.1	11	0.2	0.1	2	4	22	731	64	<LOD	2	-			
Ce	95	23	22	70	29	0.1	0.1	0.1	0.1	0.1	5	1	54	15	2	17	38	260	-	64	5	17	-			
Pr	8	0.2	0.1	12	3	<LOD	<LOD	0.02	<LOD	<LOD	1	0.1	11	0.3	0.3	4	9	33	377	115	0.1	4	-			
Nd	34	3	2	79	21	0.2	0.2	0.2	0.1	0.2	6	1	68	2	3	31	70	209	1920	950	1	27	-			
Sm	18	6	3	57	18	1	1	1	1	1	5	2	47	2	6	26	50	90	1170	153	2	25	3440			
Eu	4	1	1	15	4	0.2	0.2	0.2	0.2	0.2	2	0.2	9	0.5	0.1	1	2	5	7	2	0.1	1	18			

Gd	57	30	15	131	54	12	10	12	12	10	24	14	144	12	23	49	70	106	450	89	13	75	1380
Tb	18	10	5	33	16	7	6	8	8	7	11	7	48	4	6	12	14	20	40	13	5	19	115
Dy	201	128	58	327	182	139	111	143	143	127	177	98	489	54	58	106	111	158	196	97	76	168	457
Ho	69	48	21	94	62	64	52	65	66	59	73	38	120	20	15	28	26	39	38	24	29	45	71
Er	309	230	105	391	288	388	320	394	407	354	419	188	369	101	55	104	91	136	124	89	131	180	200
Tm	63	49	23	77	62	112	94	112	118	103	115	44	55	23	10	19	16	24	21	17	26	35	31
Yb	534	439	215	647	552	1305	1090	1284	1368	1193	1307	434	381	225	80	146	118	178	157	132	214	296	231
Lu	104	92	48	122	116	297	248	290	309	269	295	91	64	48	14	25	20	30	27	23	39	58	38
LRE	212	32	27	232	73	1	1	1	1	1	19	4	190	19	11	80	170	614	4198	1346	8	75	3440
HR	1357	1026	490	1837	1334	2324	1930	2308	2430	2122	2423	915	1679	488	261	490	468	696	1058	486	533	877	2540
EE	1569	1059	517	2069	1407	2325	1931	2309	2432	2124	2442	919	1869	507	272	570	638	1310	5256	1832	542	952	5980
REE	1569	1059	517	2069	1407	2325	1931	2309	2432	2124	2442	919	1869	507	272	570	638	1310	5256	1832	542	952	5980
Yb/ Gd <sub>N</sub>	11.64	18.19	17.93	6.11	12.73	138.1 4	140.3 8	127.7 1	144.8 1	141.1 5	67.50	38.07	3.27	23.07	4.25	3.65	2.08	2.08	0.43	1.83	19.91	4.89	0.21
Ce/ Yb <sub>N</sub>	0.05	0.01	0.03	0.03	0.01	0.000 03	0.000 02	0.000 02	0.000 01	0.000 01	0.001	0.001	0.04	0.02	0.01	0.03	0.08	0.4	-	0.1	0.01	0.01	-
Eu/ Eu*	0.34	0.30	0.51	0.52	0.43	0.17	0.17	0.16	0.17	0.17	0.44	0.14	0.33	0.26	0.03	0.09	0.12	0.16	0.03	0.04	0.04	0.07	0.02
Ce/ Ce*	1.09	26.09	-	1.30	2.49	-	-	-	-	-	1.15	3.77	1.22	14.55	2.43	1.39	1.52	2.34	-	0.18	-	1.43	-

Table 5

Lithology	Phyllite N	Phyllite S	Metatexite					Type-1 Diatexite		Type-2 Diatexite	Granite X	Granite IX	Granite VIII	Granite V	Granite II
Sample	JTJ-2	JTJ-19	JTJ-20	JTJ-25	JTJ-48B	JTJ-53	JTJ-60A	JTJ-43	JTJ-55	JTJ-24A	JTJ-34	JTJ-45	JTJ-35	JTJ-12	JTJ-36
<sup>87</sup> Sr/ <sup>86</sup> Sr	0.7250	0.7402	0.7245	0.7321	0.7318	0.7403	0.7489	0.7364	0.7340	0.7447	0.7648	0.7893	0.7636	0.7774	0.7578
±2σ	0.000013	0.000016	0.000019	0.000016	0.000013	0.000017	0.000034	0.000018	0.000065	0.000020	0.000015	0.000024	0.000025	0.000013	0.000021
<sup>87</sup> Sr/ <sup>86</sup> Sr <sub>320</sub>	0.7153	0.7089	0.7097	0.7124	0.7039	0.7099	0.7053	0.7072	0.7204	0.7162	0.7055	0.7062	0.7126	0.7022	0.7063
<sup>143</sup> Nd/ <sup>144</sup> Nd	0.5119	0.5126	0.5122	0.5121	0.5120	0.5121	0.5121	0.5120	0.5123	0.5122	0.5120	0.5121	0.5120	0.5120	0.5123
±2σ	0.000011	0.000006	0.000009	0.000010	0.000007	0.000009	0.000009	0.000009	0.000011	0.000007	0.000009	0.000008	0.000009	0.000012	0.000012
εNd	-13.42	0.15	-8.30	-10.91	-12.28	-10.57	-10.99	-12.77	-7.16	-8.28	-12.66	-11.06	-12.18	-11.51	-7.49
εNd <sub>320</sub>	-10.07	3.28	-5.21	-7.65	-8.86	-7.30	-8.08	-8.32	-5.05	-6.21	-8.89	-7.99	-8.36	-8.64	-6.03
Nd T <sub>DM2</sub> (Ma) <sup>3</sup>	1826	779	1445	1637	1731	1609	1670	1524	1690	1433	1714	1734	1693	1664	1510
<sup>176</sup> Hf/ <sup>177</sup> Hf	0.2823	0.2824	0.2825	0.2824	0.2824	0.2825	0.2824	0.2824	0.2825	0.2824	0.2823	0.2824	0.2824	0.2824	0.2825
±2σ	0.000005	0.000005	0.000004	0.000009	0.000005	0.000004	0.000005	0.000005	0.000005	0.000007	0.000006	0.000006	0.000005	0.000004	0.000005
εHf	-18.14	-11.40	-8.98	-11.41	-12.09	-9.44	-12.82	-14.92	-9.25	-12.02	-15.16	-13.99	-14.21	-14.06	-10.16
εHf <sub>320</sub>	-13.68	-7.16	-4.45	-7.99	-8.47	-6.45	-8.16	-8.16	-3.42	-9.25	-8.43	-7.78	-7.71	-7.70	-3.99

Hf T <sub>DM2</sub> (Ma) <sup>b</sup>	2160	1747	1574	1800	1830	1701	1810	1880	1810	1508	1781	1828	1781	1786	1544
<sup>208</sup> Pb/ <sup>204</sup> Pb	38.7956	38.6236	38.6820	38.8925	39.2638	38.6856	38.4898	39.9730	38.4068	38.2375	42.1802	38.6802	39.2789	38.9877	38.3497
±2σ	0.0021	0.0023	0.0023	0.0021	0.0018	0.0013	0.0022	0.0022	0.0026	0.0024	0.0022	0.0021	0.0017	0.0014	0.0023
<sup>206</sup> Pb/ <sup>204</sup> Pb	18.3550	19.0401	18.7160	18.6505	18.6196	19.0338	18.5718	18.5110	18.4268	18.2206	19.0352	18.6156	18.6437	18.6888	18.6227
±2σ	0.0009	0.0010	0.0011	0.0010	0.0009	0.0005	0.0011	0.0010	0.0011	0.0009	0.0009	0.0010	0.0007	0.0006	0.0009
<sup>207</sup> Pb/ <sup>204</sup> Pb	15.6924	15.6746	15.6434	15.6434	15.6485	15.6848	15.6502	15.6483	15.6407	15.6246	15.6830	15.6541	15.6583	15.6631	15.6618
±2σ	0.0008	0.0008	0.0009	0.0008	0.0007	0.0005	0.0009	0.0008	0.0010	0.0008	0.0008	0.0008	0.0007	0.0005	0.0008
<sup>208</sup> Pb/ <sup>206</sup> Pb	2.1136	2.0285	2.0668	2.0853	2.1087	2.0325	2.0724	2.1594	2.0843	2.0986	2.2159	2.0778	2.1068	2.0862	2.0593
±2σ	0.000029	0.000033	0.000037	0.000046	0.000034	0.000027	0.000037	0.000041	0.000032	0.000037	0.000030	0.000046	0.000039	0.000045	0.000086
<sup>207</sup> Pb/ <sup>206</sup> Pb	0.8549	0.8232	0.8358	0.8388	0.8404	0.8241	0.8427	0.8453	0.8488	0.8575	0.8239	0.8409	0.8399	0.8381	0.8410
±2σ	0.000011	0.000011	0.000013	0.000016	0.000010	0.000008	0.000011	0.000014	0.000010	0.000010	0.000010	0.000015	0.000010	0.000015	0.000024
<sup>206</sup> Pb/ <sup>204</sup> Pb <sub>320</sub>	17.8770	17.9549	17.9149	17.9774	17.7057	12.7938	18.0033	17.7218	17.9120	17.2398	18.1733	17.4132	17.5719	17.6702	17.9340
<sup>207</sup> Pb/ <sup>204</sup> Pb <sub>320</sub>	15.6671	15.6173	15.6011	15.6079	15.6002	15.3554	15.6201	15.6067	15.6135	15.5728	15.6375	15.5907	15.6017	15.6093	15.6254
<sup>208</sup> Pb/ <sup>204</sup> Pb <sub>320</sub>	38.0388	37.6798	38.1323	38.2769	38.6531	38.1212	38.2232	36.6712	38.2117	38.2151	39.3876	38.1326	37.5695	38.2999	38.1579

$${}^a T_{DM2} = 1/\lambda \times \ln\left(\frac{{}^{143}\text{Nd}/{}^{144}\text{Nd} - (e^{-\lambda t} - 1) \times ({}^{147}\text{Sm}/{}^{144}\text{Nd} - {}^{147}\text{Sm}/{}^{144}\text{Nd}_{CC}) - {}^{143}\text{Nd}/{}^{144}\text{Nd}_{DM}}{({}^{147}\text{Sm}/{}^{144}\text{Nd}_{CC} - {}^{147}\text{Sm}/{}^{144}\text{Nd}_{DM}) + 1}\right)$$

$${}^b T_{DM2} = 1/\lambda \times \ln\left(\frac{{}^{176}\text{Hf}/{}^{177}\text{Hf} - (e^{-\lambda t} - 1) \times ({}^{176}\text{Lu}/{}^{177}\text{Hf} - {}^{176}\text{Lu}/{}^{177}\text{Hf}_{CC}) - {}^{176}\text{Hf}/{}^{177}\text{Hf}_{DM}}{({}^{176}\text{Lu}/{}^{177}\text{Hf}_{CC} - {}^{176}\text{Lu}/{}^{177}\text{Hf}_{DM}) + 1}\right)$$

λRb - Steiger and Jäger (1977)

λSm - Lugmair and Marti (1978)

λLu - Söderlund et al. (2004)

λU and λTh - Steiger and Jäger (1977)

<sup>147</sup>Sm/<sup>144</sup>Nd<sub>CHUR</sub> and <sup>143</sup>Nd/<sup>144</sup>Nd<sub>CHUR</sub> - Jacobsen and Wasserburg (1980)

<sup>147</sup>Sm/<sup>144</sup>Nd<sub>DM</sub>, <sup>143</sup>Nd/<sup>144</sup>Nd<sub>DM</sub>, and <sup>147</sup>Sm/<sup>144</sup>Nd<sub>CC</sub> - Liew and Hofmann (1988)

<sup>176</sup>Lu/<sup>177</sup>Hf<sub>CHUR</sub> and <sup>176</sup>Hf/<sup>177</sup>Hf<sub>CHUR</sub> - Blichert-Toft and Albarède (1997)

<sup>176</sup>Lu/<sup>177</sup>Hf<sub>DM</sub> and <sup>176</sup>Hf/<sup>177</sup>Hf<sub>DM</sub> - Vervoort and Blichert-Toft (1999)

<sup>176</sup>Lu/<sup>177</sup>Hf<sub>CC</sub> - Griffin et al. (2002)



Fig.1

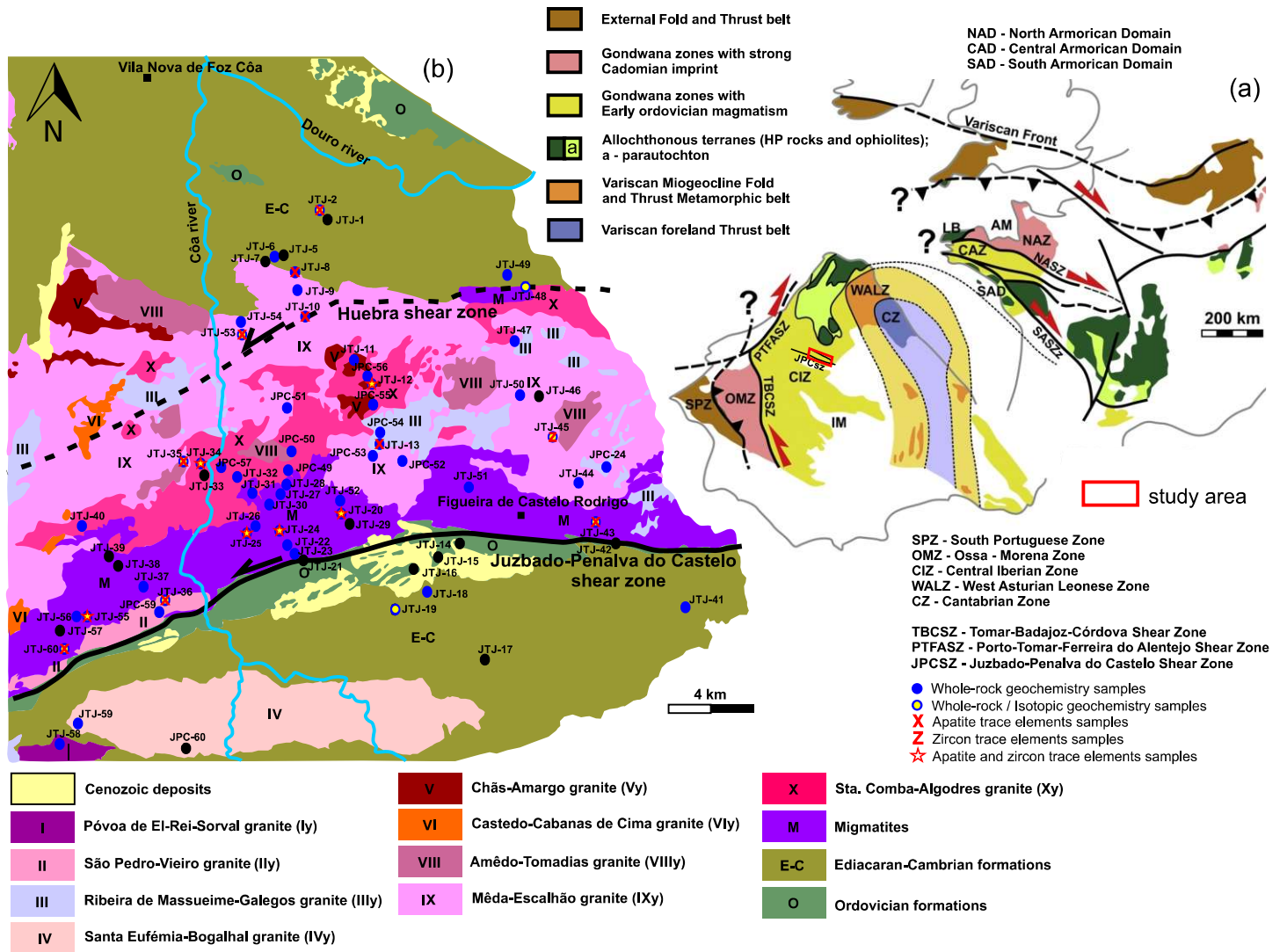




Fig.2

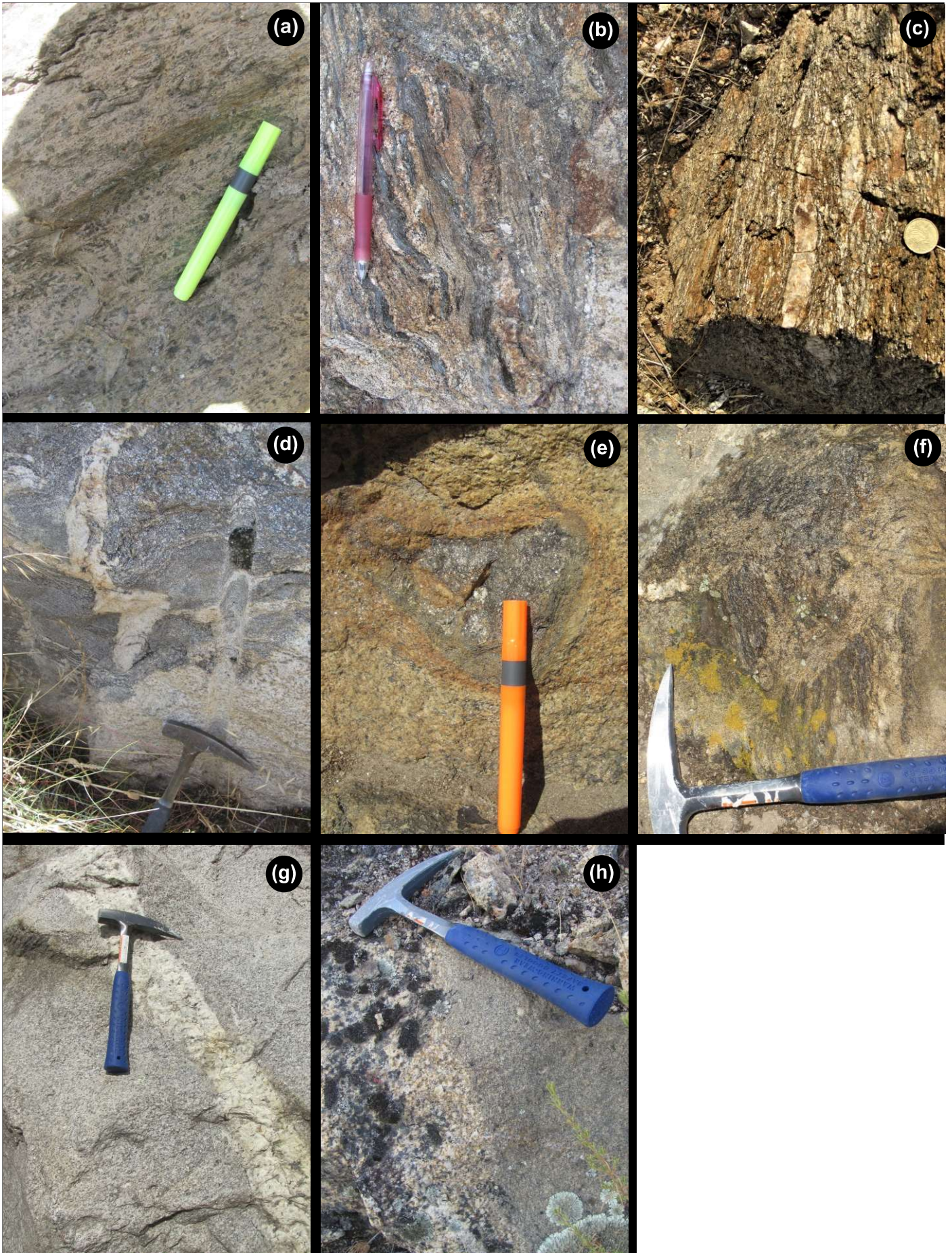
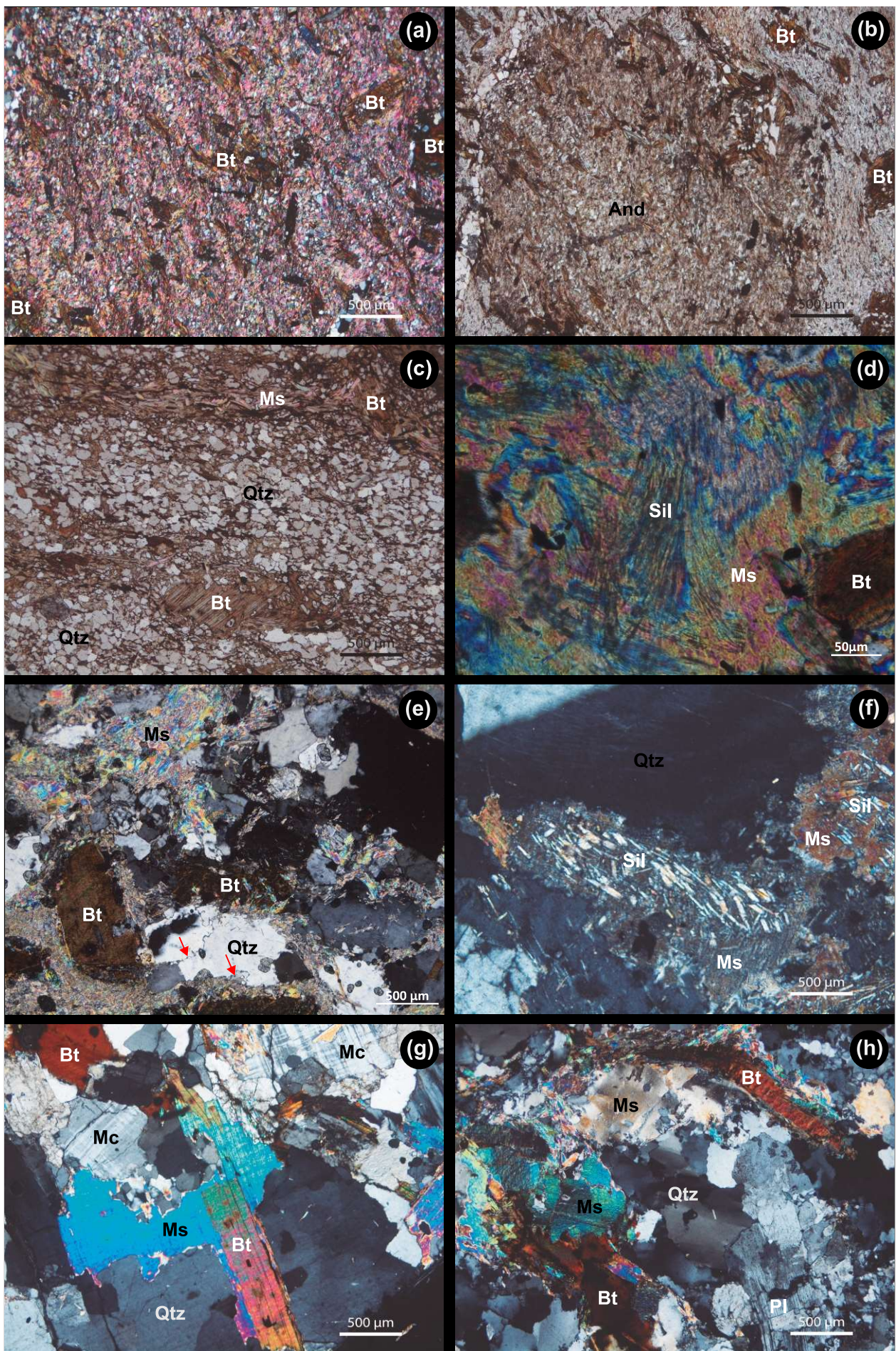
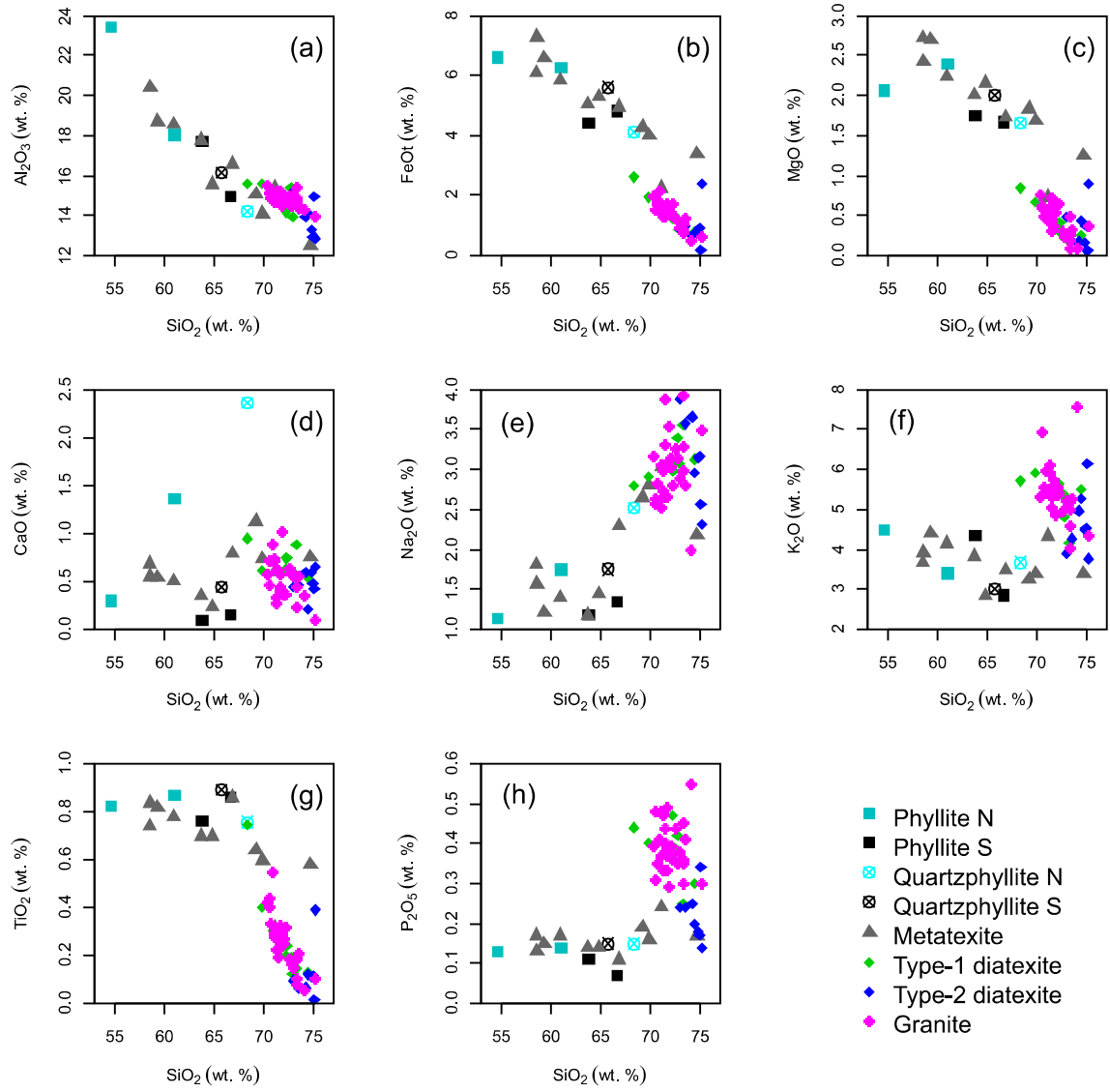




Fig.3





**Fig.4**

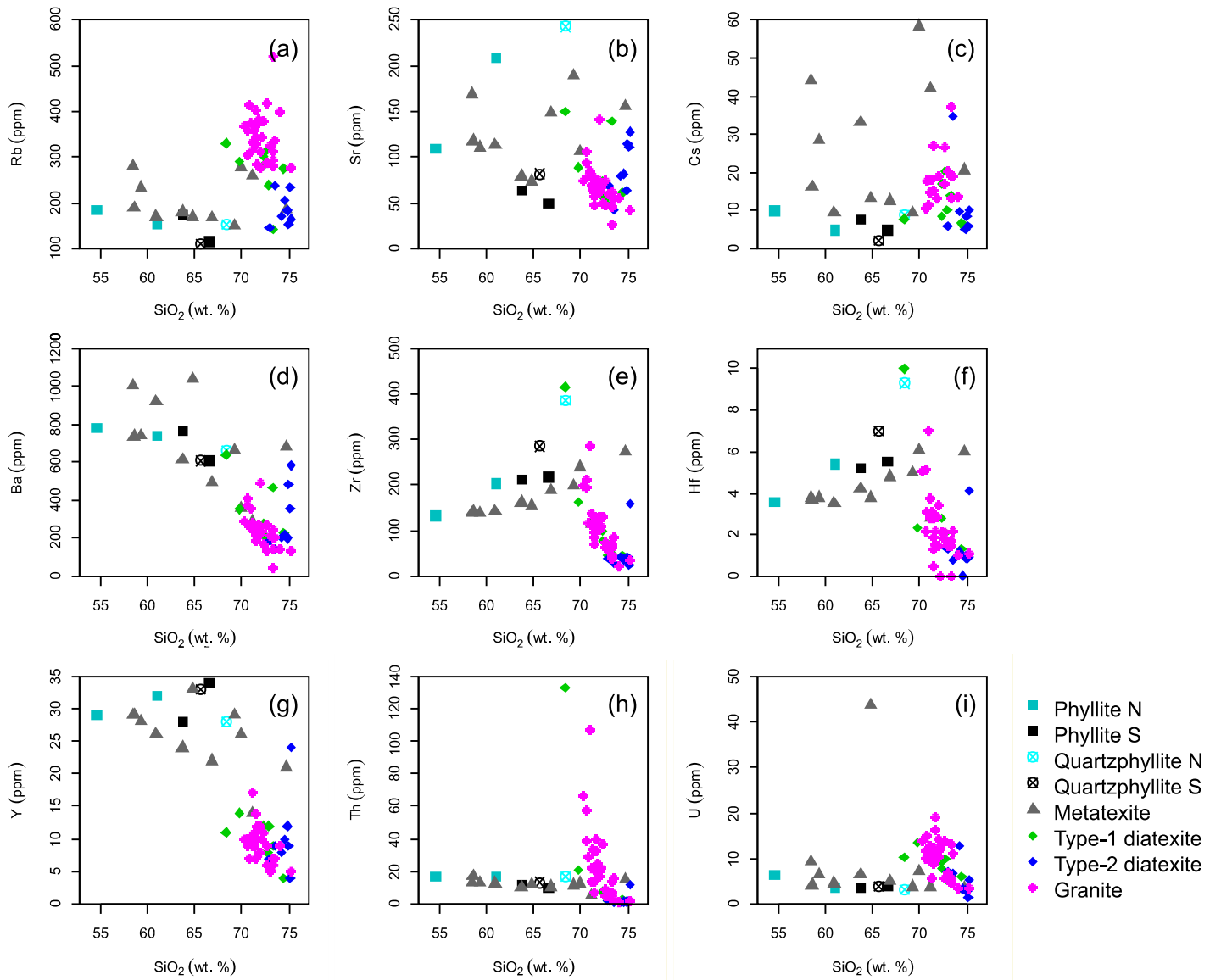
**Fig.5**

Fig.6

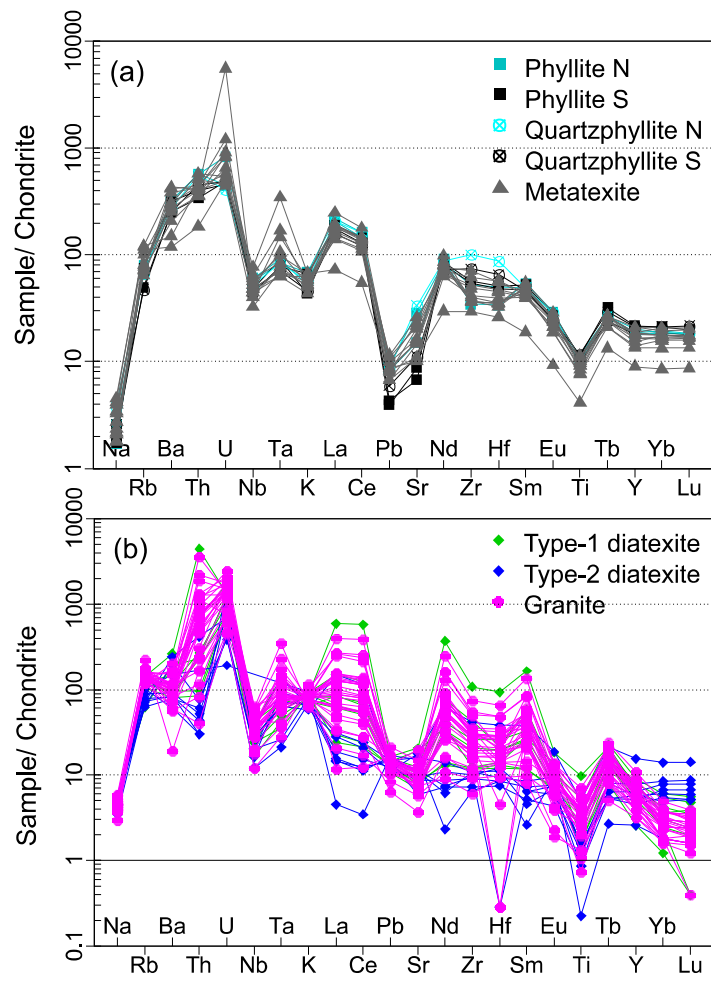
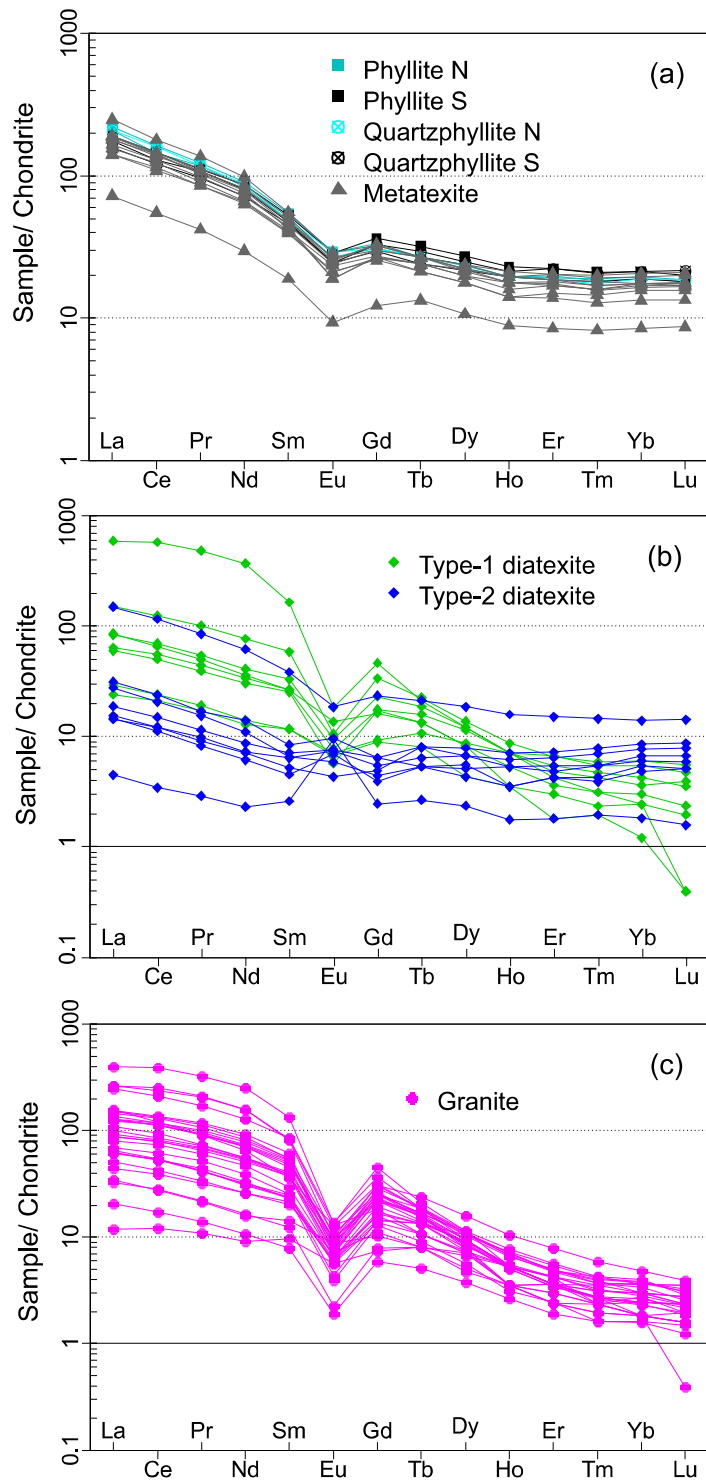
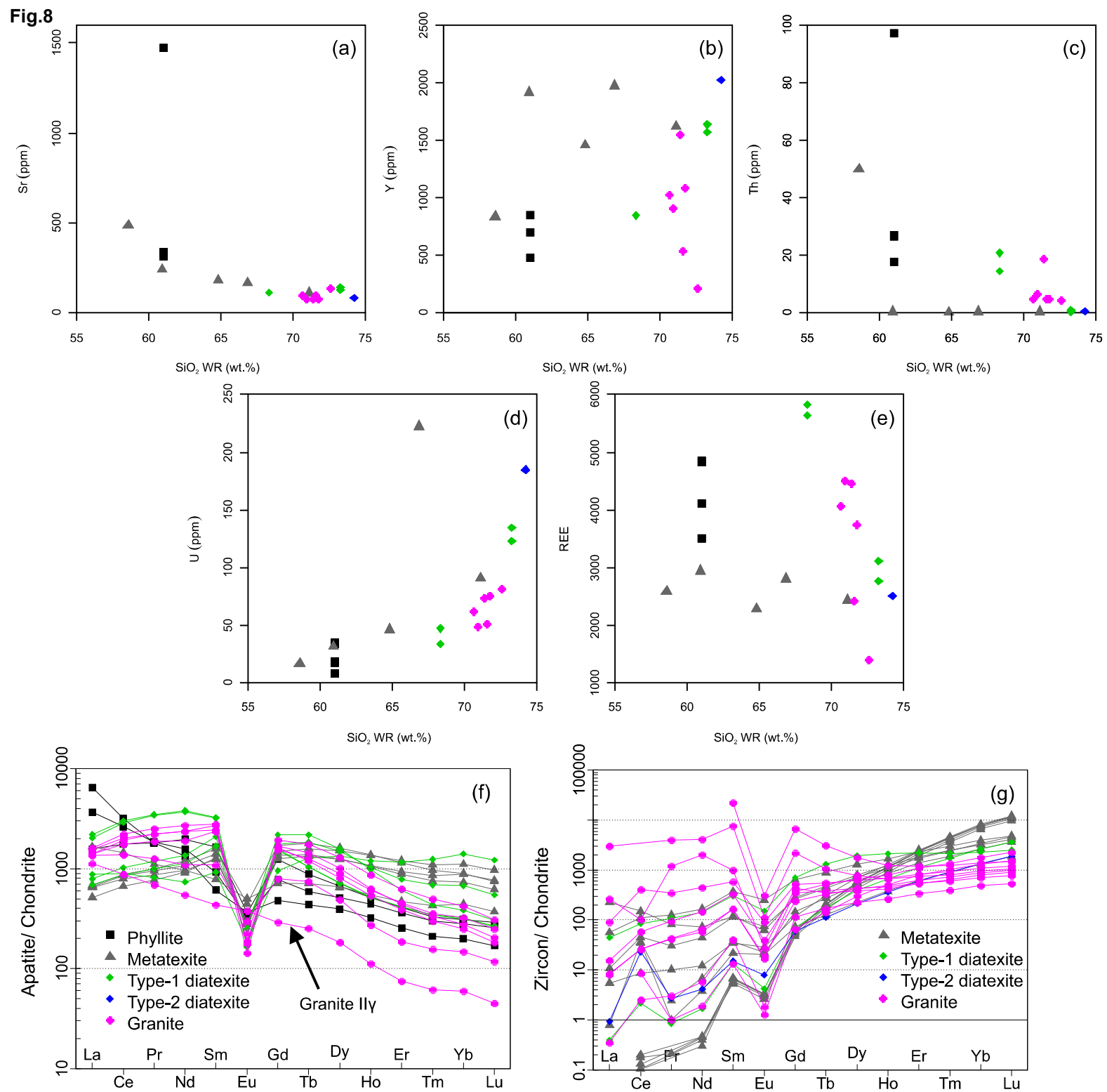
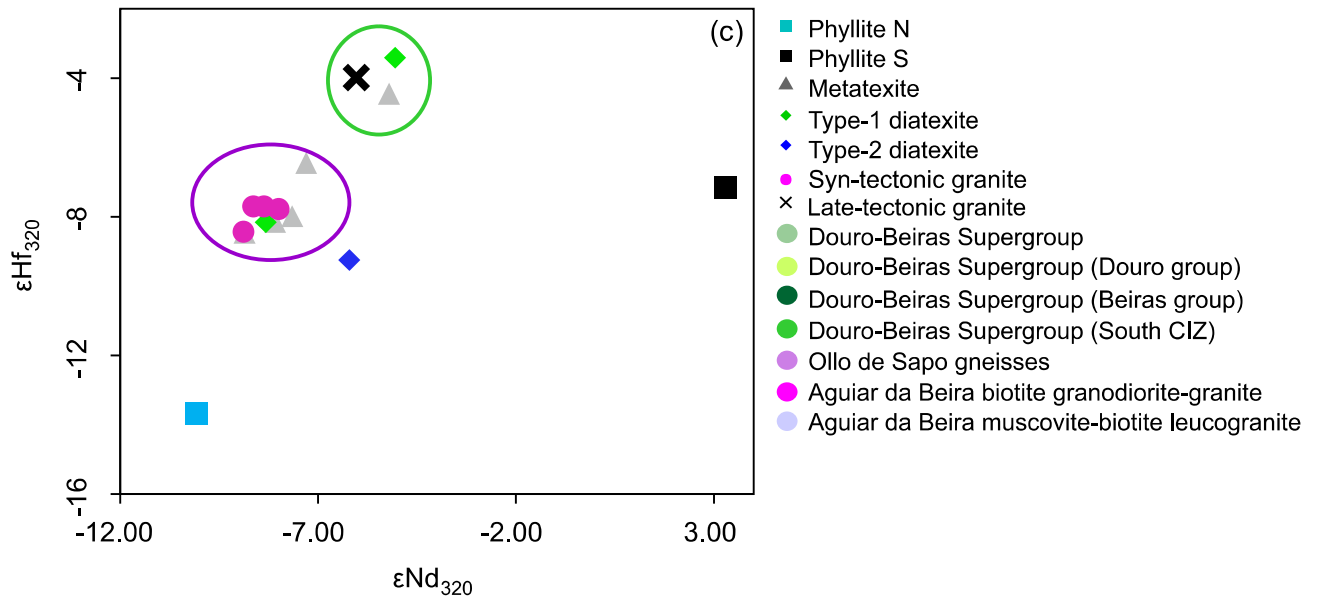
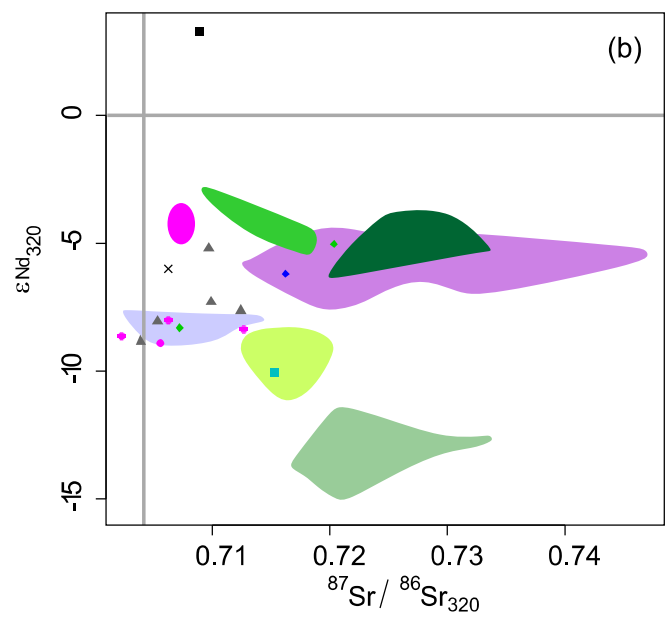
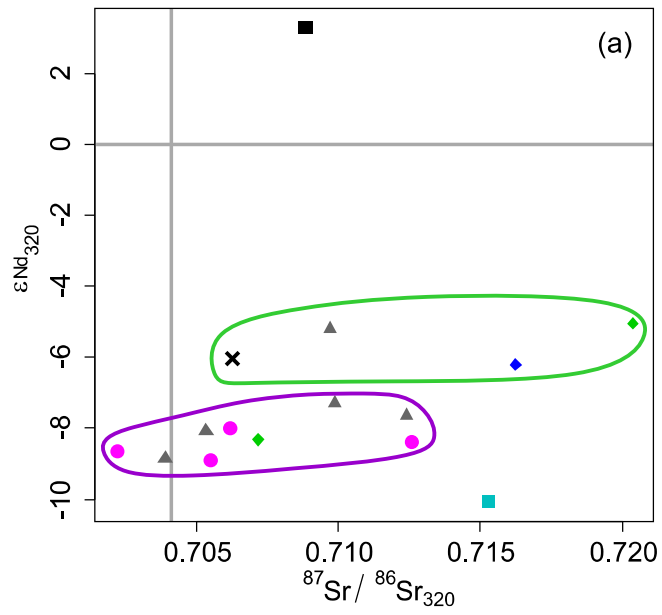


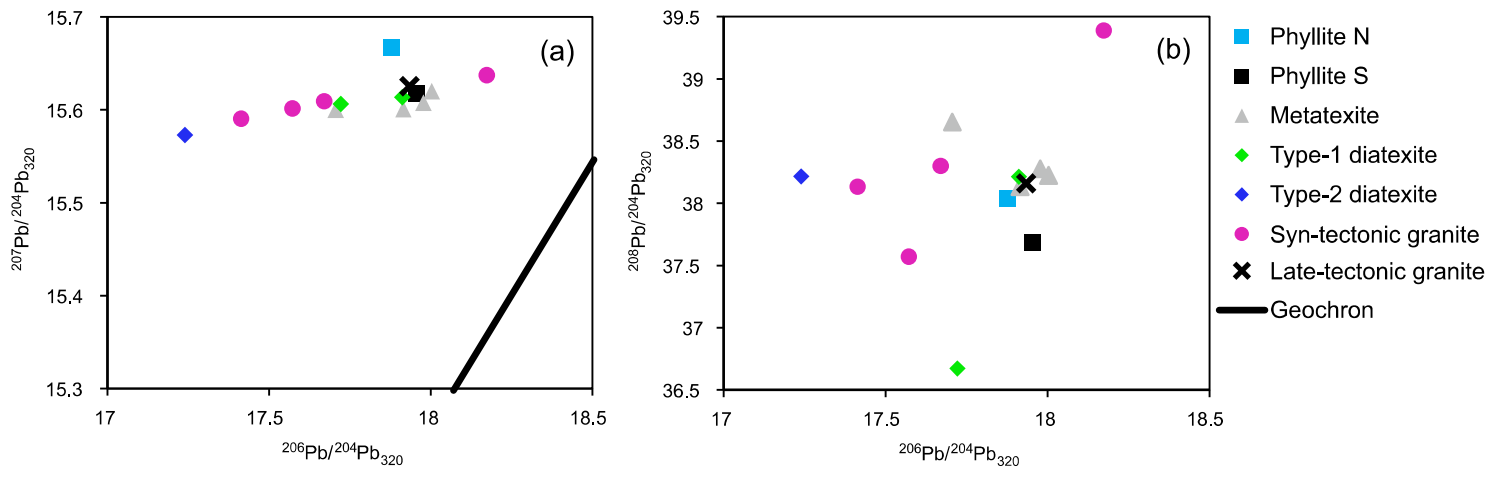
Fig.7

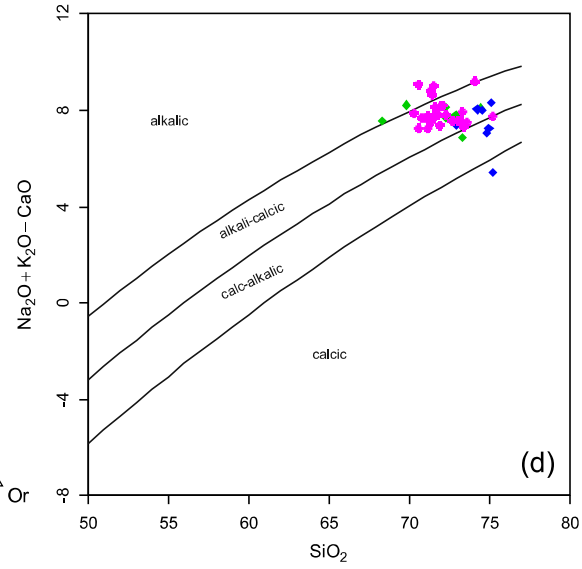
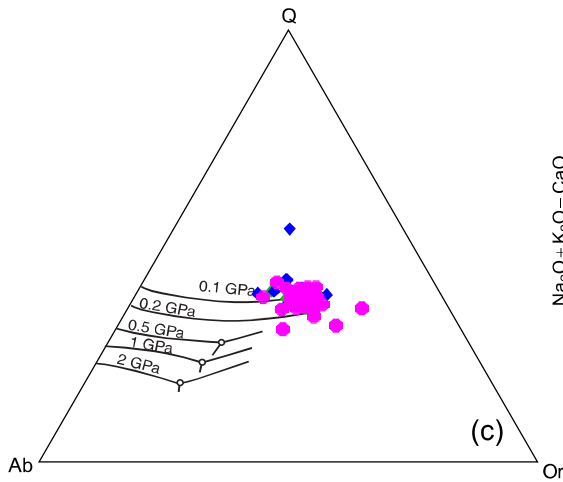
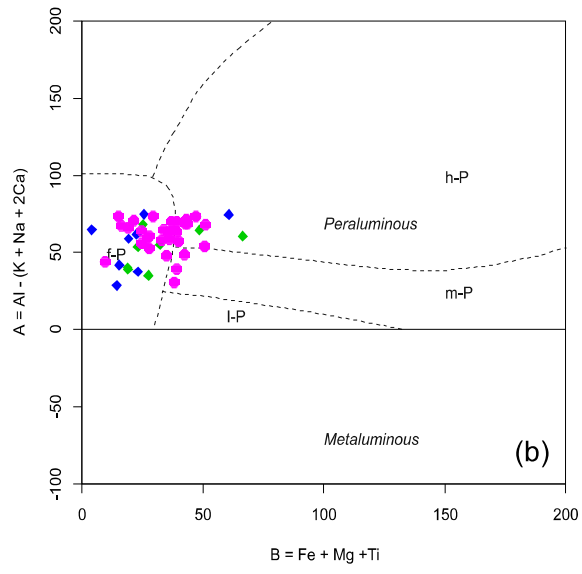
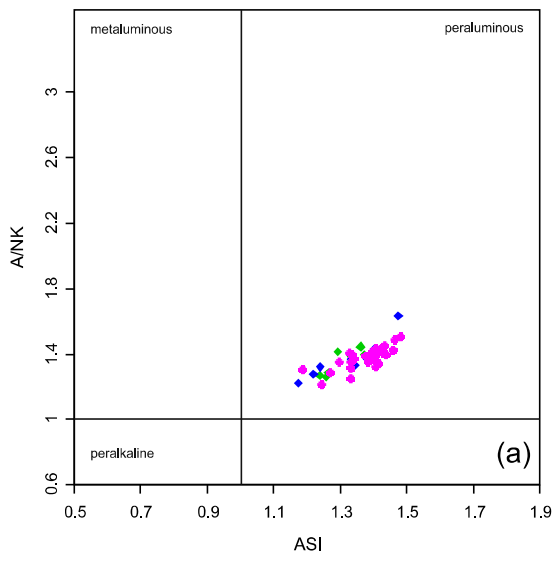


**Fig.8**



**Fig.9**

**Fig.10**

**Fig.11**

- Phyllite N
- Phyllite S
- ⊠ Quartzphyllite N
- ⊠ Quartzphyllite S
- ▲ Metatexite
- ◆ Type-1 diatexite
- ◆ Type-2 diatexite
- Granite

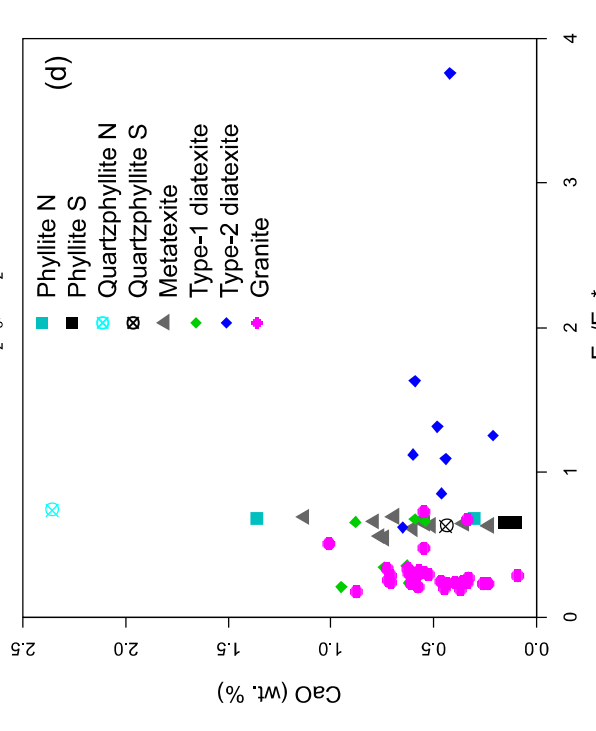
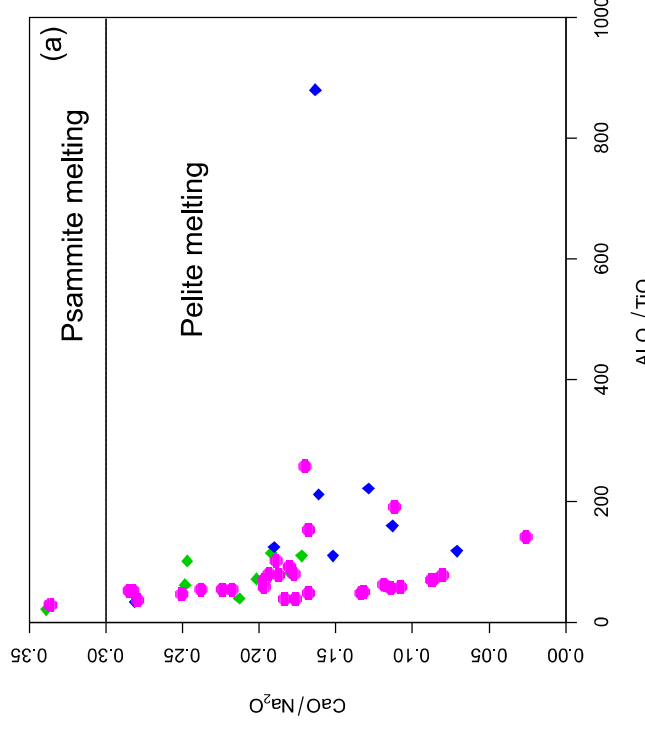
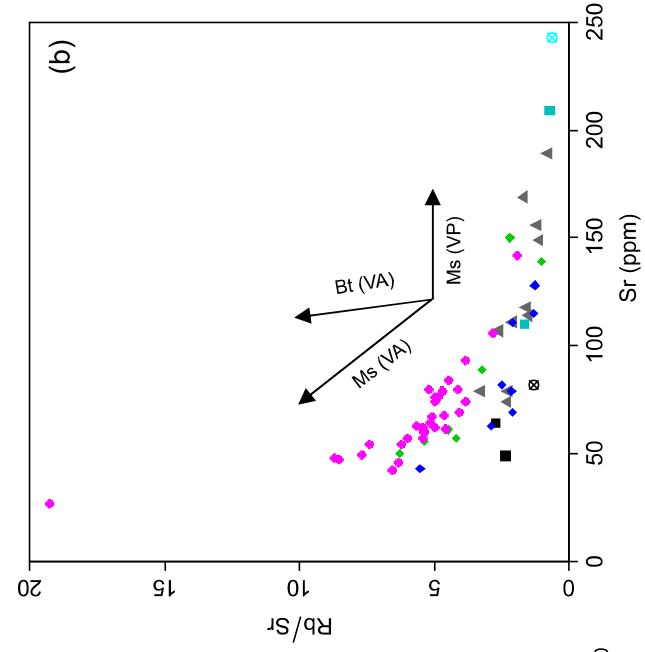
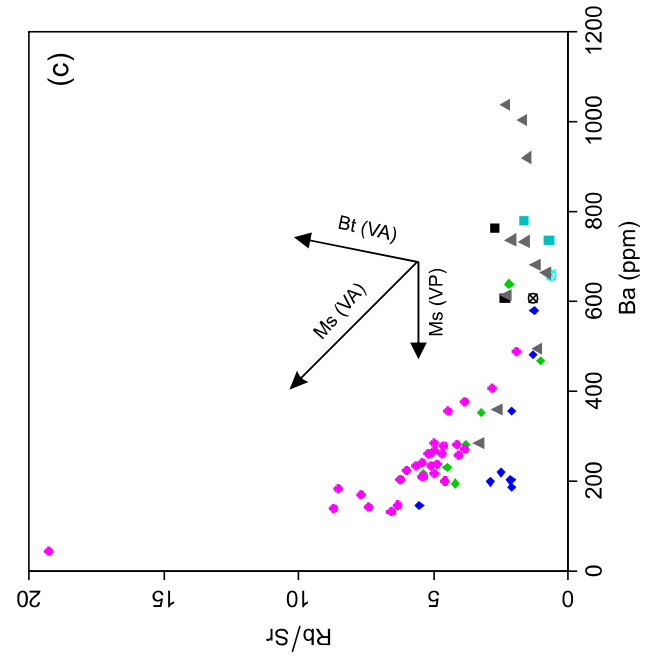


Fig.12

Fig.13

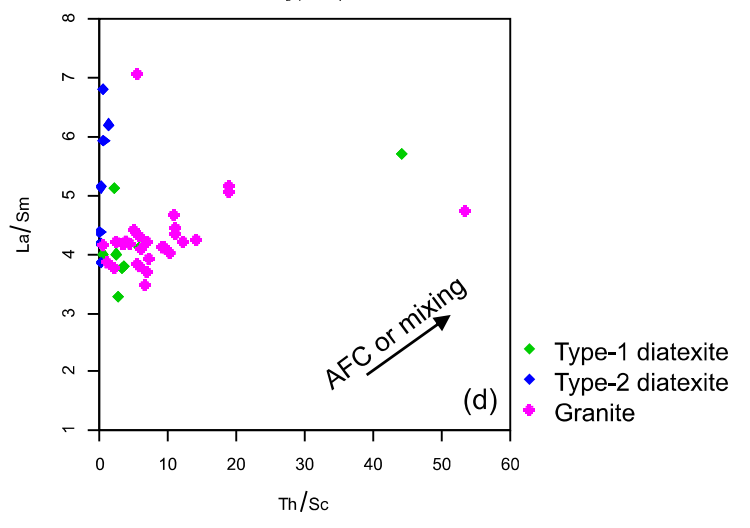
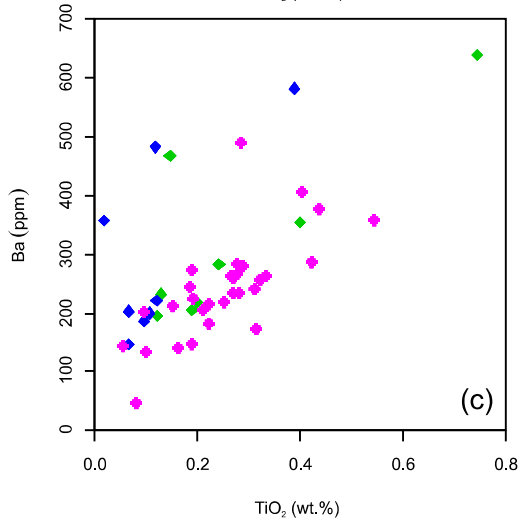
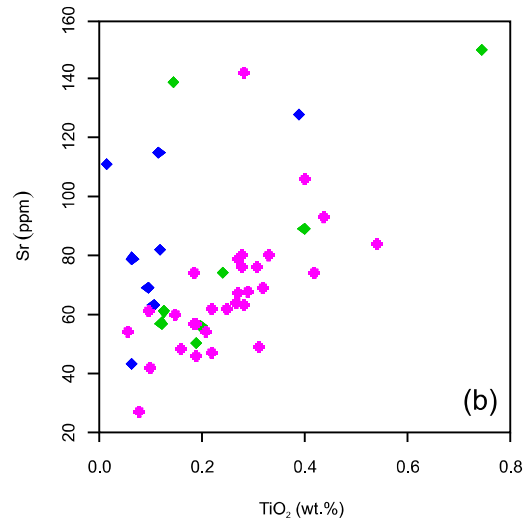
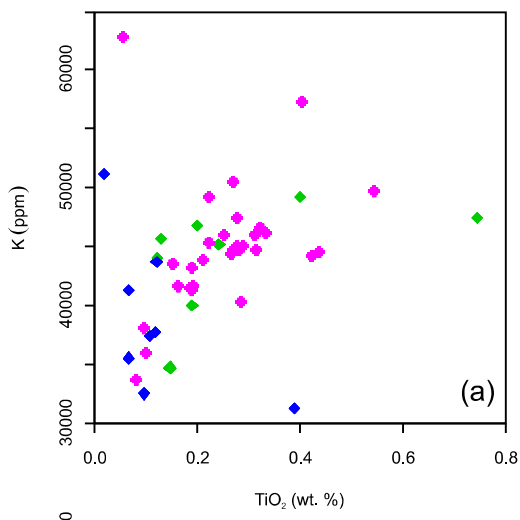
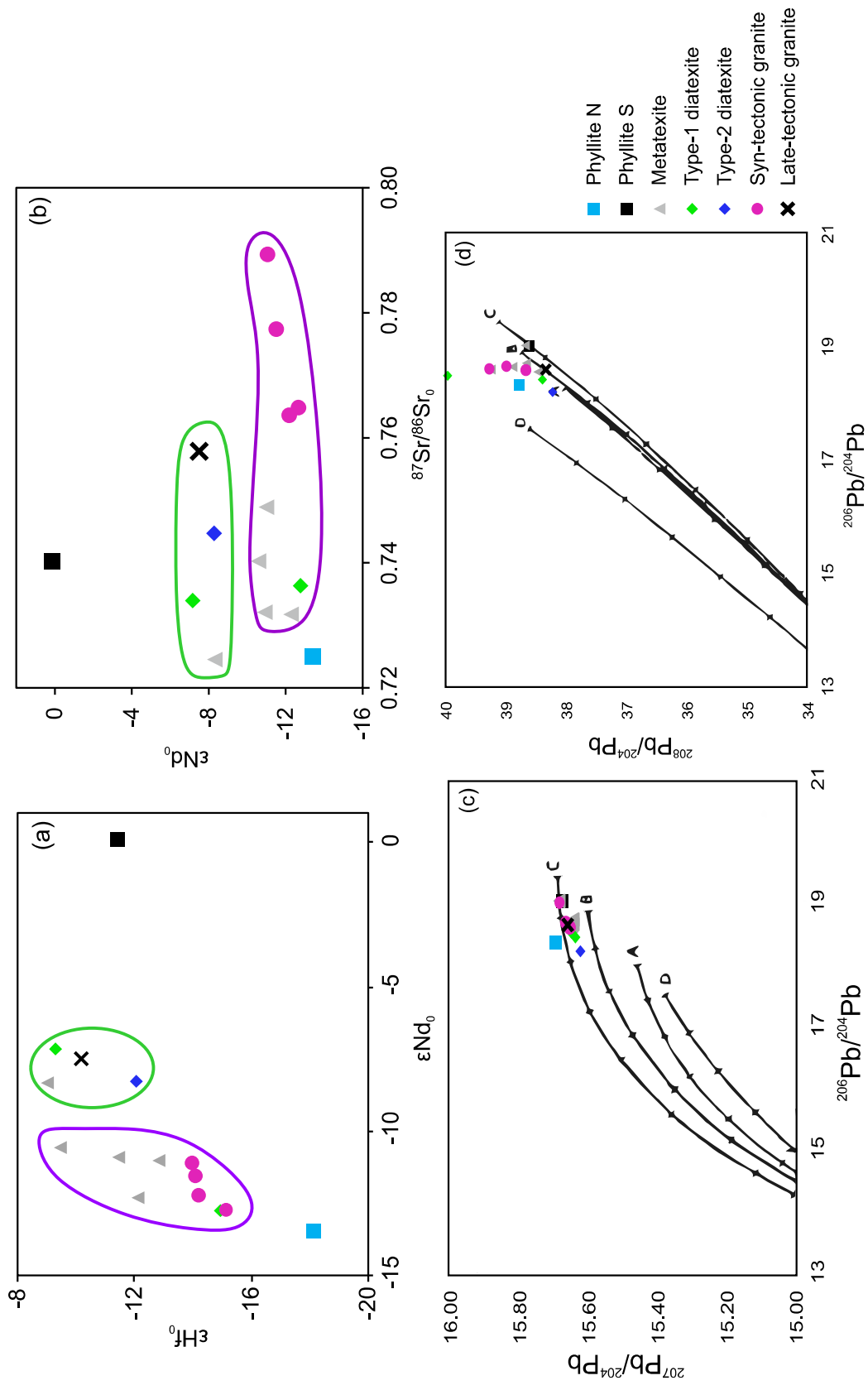
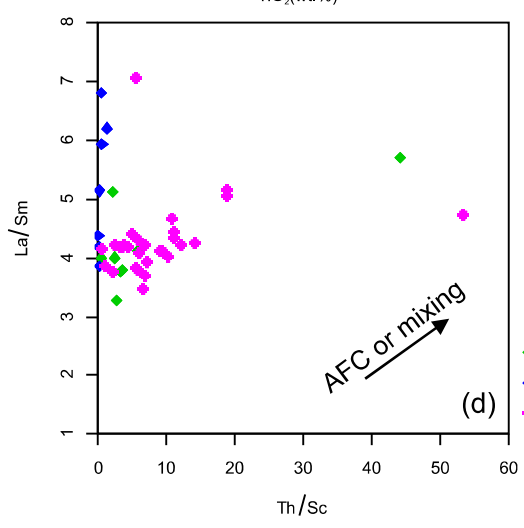
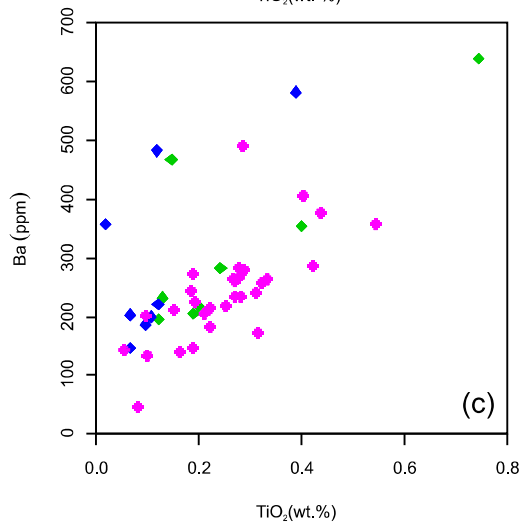
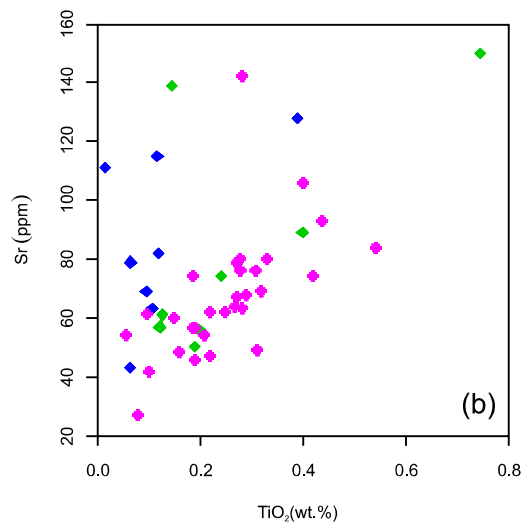
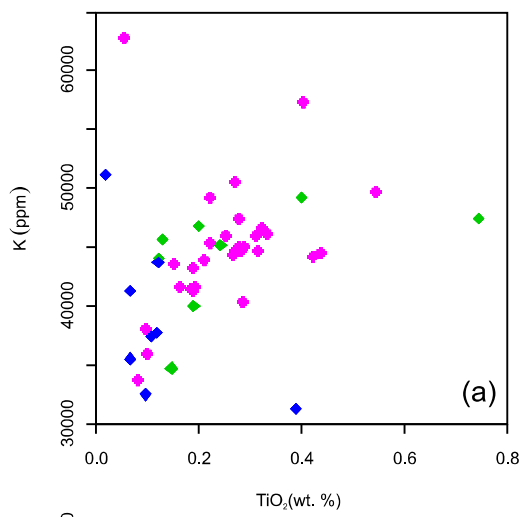


Fig.14



**Fig.15**



- ◆ Type-1 diatexite
- ◆ Type-2 diatexite
- ◆ Granite